Barrier evolution of Cape San Blas, Saint Joseph Peninsula, Florida from textural analysis, ground penetrating radar and organic matter isotope geochemistry

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Sea-Level and Barrier evolution of St. Joseph Peninsula, North Gulf of Mexico, Florida from textural analysis, ground penetrating radar and organic matter isotope geochemistry.

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ABSTRACT

St. Joseph peninsula is situated on the panhandle of Florida west coast in the northeastern Gulf of Mexico at N29°50’ and W85°20’ and is located at the west edge of the westernmost portion of the Apalachicola Barrier Island Complex (ABIC) on the Gulf of Mexico shoreline. Three vibra-cores were collected on Saint Joseph Bay side of Cape San Blas which is part of St. Joseph peninsula to determine its evolution in context of previous work by Rink and Lopez (2010). The study uses detailed textural analysis (PSD - Particle Size Distribution plots), multivariate statistics on the PSDs (Q-mode cluster analysis) and organic matter geochemistry (C/N and $\delta^{13}C$). In addition, Ground Penetrating Radar (GPR) profiles are used to provide broader stratigraphic context.

The stratigraphic analysis found that CSB has an older nucleus of strandplain deposits dating to >12 Ka that were subsequently flooded and modified through Holocene sea-level rise at ≈ 2.2 Ka. Actual barrier formation began sometime between 2.2. Ka and 0.6 Ka which is the oldest beach ridge measured by Rink and Lopez (2010).

Progradation of the barrier on the St Joseph Bay side began at least by 0.3 Ka and likely earlier. There is no evidence to indicate a higher than present sea-level in our core data and our data follows that of other sea-level studies using submerged offshore samples.

1. INTRODUCTION

Traditionally, foraminiferal and thecamoebian analysis is used for paleo-environmental interpretation and reconstructing sea level changes in coastal areas such as marshes and wetlands (Horton, 1999; Gehrels et al., 2001; Patterson et al., 2005). However, in some instances foraminifera are not present or are not preserved for a paleo-environmental determination (i.e. fresh vs. marine peat; Shennan et al., 1999; Gonzalez et al., 2000; Wilson et al., 2005b) making sea-level reconstruction difficult. In St. Joseph's Peninsula, there are no foraminifera in the sandy beach sediments or in the fringing marshes and there are only sparse thecamoebian populations in the wetlands. To address this shortcoming, this study uses detailed textural analysis (PSD - Particle Size Distribution plots; Donato et al., 2009) and organic matter geochemistry (C/N and $\delta^{13}C$) of three vibra-cores taken from the lee side of the barrier. In addition, Ground Penetrating Radar (GPR) profiles are used to provide broader stratigraphic context. The goal of this study is to evaluate the formation and evolution of part of the St. Joseph peninsula (the western most portion of Apalachicola Barrier Island Complex; Fig. 1) and to compare the data with previous results (Stapor, 1975; Forrest, 2003; Otvos, 2005a; Rink and López Rink, 2010).

1.1. Sea-Level Change and Barrier Formation

Research has examined the response of the northern coast of the Gulf of Mexico to differing rates of sea-level change during the Holocene (Penland et al., 1991; Thomas and Anderson, 1994). Differences in the sea level curves have created extensive debate (e.g. gradual rise, stillstand, timing etc.; Shepard, 1964; Kidson, 1982; Pirazzoli, 1991) with the mid-Holocene highstand being a focus of attention (Donoghue et al., 1998; Otvos, 1999). Reconstructions based on data from the Mississippi delta and surrounding areas indicate steady sea level rise during the Holocene (Tornqvist et al., 2006; Write et al., 2005; Milliken et al., 2008). This is counter to Coleman and Smith (1964) who proposed that regional sea level reached its present level by ca. 2-5 ka, and load-induced subsidence was responsible for late Holocene sea level rise in the delta region. Tornqvist et al. (2004) argues it is due to glacio-isostatic adjustments. Research in other coastal environments (barriers, lagoons etc.) of the Gulf (Morton et al., 2000; Blum et al., 2001, 2002, and 2003) favor a mid-Holocene highstand - i.e. sea level was 1-2m higher than present. However, this is not universally accepted (e.g. see Otvos, 2001; Rodriguez et al., 2004), with the cause of a higher than present sea level proving uncertain (Stapor et al., 1991; Tanner, 1992; Morton et al., 2000). Many contradictions among the data sets have prevented consensus among researchers (Kidson, 1982). Behrens (1966a) presented a date (i.e. ca. 2000 ^{14}C years B.P.) for a beach deposit, located at 4 m higher than the present sea level, at a location close to the US-Mexican border. Additional research on other prograding beach ridges along the Florida coast in the eastern Gulf of Mexico provided evidence for two mid-Holocene highstands (Tanner et al., 1989). However, recent research on coastal marshes in Florida did not provide any support for this (Write et al., 2005). Most of the evidence for the mid-Holocene highstand (+2 m at ~ 5 ka years BP) comes from studies on coastal land forms (i.e. beach ridges, marshes, subtidal flats and wave cut features) in Texas (Morton et al., 2000; Blum et al., 2001). The mid-Holocene opponents (Otvos, 2001, 2004; Rodriguez et al., 2004) believe the elevated beach ridges are formed as a result of more intense storm activity during that time (Rodriguez and Meyer, 2006; Donnelly and Giosan, 2008).

About 15% of the world's shorelines have barrier islands and are a prominent feature of the Atlantic and Gulf Coasts of the United States (Fisher, 1982). Over the years, there has been significant debate on their origin (e.g. Redman, 1952; Johnson, 1919; Davies, 1961; Bigarella, 1965; Currey et al., 1967; Psuty, 1967; Alexander, 1969; Tanner, 1970; Wright, 1970; Stapor, 1975; Carter, 1986; Mason, 1991) and there have been many studies carried out worldwide (Johnson, 1919; Alexander, 1969; Carter, 1986, Davis, 1994; Pekala, 1996; Locker et al., 1998, 2003). However, their origin is generally thought to be due to a combination of processes (Schwartz, 1971).

De Beaumont (1845) was among the first to propose the concept of bar emergence as a mechanism for barrier island formation. This concept is favored in studies from the eastern and northern coasts of Gulf of Mexico (Otvos, 1970, 1979, 1981, 1984) and is where the concept of bar emergence was further developed (Davis, 1994; Davis and Hine, 1989; Gibbs and Davis, 1991). Gilbert (1885) proposed spit breaching and development by longshore drift as another mechanism for barrier formation and evolution on middle Atlantic and Gulf of Mexico coasts (Fisher, 1968; Leatherman, 1982;

Randazzo and Jones, 1997). McGee, (1890) proposed that barriers formed with sea level rise and drowning of coastal land forms (i.e. beach ridges and dunes) parallel to the shores.

Elevations of coastal landforms (i.e. beach ridges) found on many modern barriers, have been used as sea-level indicators (e.g. Otvos, 2000; Hampson, 2000). However, the use of beach ridges as sea-level indicators depends on the understanding the dominant processes controlling their formation and preservation (Otvos, 2000). Wave energy is considered important in beach ridge development with variations in sea level, tides and wind affecting the fluctuations in wave energy and their elevation above sea level. Tanner (1995) suggested that the various characteristics (i.e. granulometry and sedimentary structures) of beach ridges and swales are the indicative of normal fair-weather wave processes. A small fall in sea level will generate swales whereas beach ridges are formed as a result of rise in the intensity of these processes by a small rise in sea level (Tanner, 1995). Otvos (2000) proposed that the combination of foreshore and aeolian processes are responsible for the development of beach ridges.

1.2. Sea-Level Indicators

1.2.1. Microfossils

Microfossils studies and in particular foraminifera and thecamoebians have been used extensively in sea-level reconstruction particularly in marine marsh peat (e.g. Scott and Medioli, 1978; Gehrels, 1994, 2000; Horton and Edwards, 2006). Foraminifera/thecamoebians are often zoned in marine marshes based on tidal flooding with distinctive assemblages living in the high and low marsh (Scott and Medioli, 1980, 1986; Scott and Leckie, 1990). This zonation is used to reconstruct sea-level often at a very high resolution (e.g. Shennan, 1986, 1992; Van de Plassche, 1986; Scott et al., 2001). However, instances where microfossils are absent due to taphonomic bias (i.e. chemical and mechanical breakdown) or are not present at all, their absence can be problematic for determining sea-level (e.g. Shennan et al., 1999; Gonzalez et al., 2000; Wilson et al., 2005b). This is particularly problematic for distinguishing between marine or freshwater peat which has important implications for sea-level.

1.2.2. Organic Matter Geochemistry

Carbon isotopes ($\delta^{13}C$) and organic carbon to total nitrogen ratios (C:N) of coastal sediments are also used as a sea-level proxy for determining sub-tidal, inter-tidal and supra-tidal deposited sediments (e.g. Shennan et al., 2000, Wilson et al., 2005; Lamb et al., 2006, 2007). The sources (terrestrial vs. marine) of carbon input to coastal and marine environments can be identified by using the compositional isotopic variations of organic carbon (Megen et al., 2002) but further discrimination is possible by including the organic carbon to total nitrogen ratio (C:N). About 90% of all terrestrial plants use the C3 photosynthetic pathway and have $\delta^{13}C$ values between -32‰ and -21‰ whereas the terrestrial plants that use the C4 photosynthetic pathway have $\delta^{13}C$ values between -17‰ to -9‰ (Deines, 1980). Terrestrial plants are predominantly composed of lignin and cellulose (nitrogen poor) and have high C/N ratios >12 (Prahl et al., 1980). Marine $\delta^{13}C$ particulate organic carbon (POC) values range between -21‰ and -18‰ (Peters et al., 1978; Wada et al., 1987; Middelburg and Nieuwenhuize, 1998) whereas C/N ratios range between 5 and 7 (Meyers, 1994; Tyson, 1995). The marine $\delta^{13}C$ dissolved organic carbon (DOC) values range between -22‰ and -25‰, whereas the freshwater $\delta^{13}C$ DOC values range between -26‰ and 28‰. In coastal environments, the two main sources of organic material in the coastal sediments have been considered as autochthonous sources (in situ growth of plants) and allochthonous sources carried by wind or water, either from the terrestrial catchment or from marine environment. The $\delta^{13}C$ and C/N ratios of sediments are a function of these sources of organic material (either in situ or transported material) and their corresponding proportions (Fry and Sherr, 1989). The $\delta^{13}C$ values of organic matter can reveal information on the source of organic carbon in an environment (Nissenbaum & Kaplan, 1972). The range of $\delta^{13}C$ values closer to that of marine algae than to that of terrestrial plant materials reveals that the sources of organic carbon are nearshore marine waters and sediments (Eadie & Jeffrey, 1973). The major autochthonous sources of organic matter in marshes are vascular plants whereas the organic material in marsh sediments (as an allochthonous source) is derived from fresh, brackish or marine environments in the form of particulate or dissolved material (Lamb et al., 2006).

1.2.3. Particle-Size Analysis

Since 1960, a large number of studies have been conducted attempting to determine the significance of textural characteristics (mean, mode, standard deviation etc.) for interpreting environmental settings (tidal flats, marshes and dunes etc.) using a graphic calculation method (Folk and Ward, 1957) or calculated moment measures (Friedman, 1961). However, subsequent research indicated that texture was not useful as an indicator of the depositional environments (Shepard and Young, 1961; Schlee et al., 1964; Moiola and Weiser, 1968; Solohub and Klovan, 1970). Recently three dimensional plots (surface plot of Particle Size Distributions PSD; Beierle et al., 2002) have been used as an alternative method of graphically representing PSDs which have been facilitated with the advent of laser particle-size characterization (e.g. Donato et al., 2009; van Hengstum et al., 2007; Reinhardt et al., 2010). Textural analysis can contribute a significant role in categorizing the particles for interpreting the sedimentary environments (Pye and Blott, 2004). The pattern of PSDs and their change with environment is a potential facies tool that has not been fully explored (e.g. Reinhardt et al., 2010).

1.2.4. Ground Penetrating Radar

Ground Penetrating Radar (GPR) has been employed in coastal environmental studies to determine large scale stratigraphic patterns (Neal et al., 2002; Barboza et al., 2009; Bennett et al., 2009; Gomez-Ortiz et al., 2009; Neilson et al., 2009). Using high frequency (MHz) electromagnetic pulses (EM), the GPR is able to acquire high resolution images of subsurface stratigraphy and depositional geometry in suitable substrates through the measurement of two- way travel time of reflected EM waves (Neal, 2004). The high frequency (MHz) electromagnetic (EM) waves from GPR reflect from sediment boundaries of contrasting electrical properties (dielectric permittivity) and provide high resolution visualization of subsurface geometry (Neal, 2004). However, there can be problems with the GPR technique in coastal environments studies where high electrical conductivity of salt water (due to an abundance of dissolved ions) acts as a constraining factor for maximum depth penetration due to rapid diffusion of electrical charges of propagating electromagnetic (EM) waves. Data collected from beach and aeolian systems has been essential in the development and refinement of coastal depositional models (Neal et al., 2002; Barboza et al., 2009; Bennett et al., 2009; Gomez-Ortiz et al., 2009; Neilson et al., 2009). The cross-sectional profiles often highlight the interface between overlying aeolian sediments and underlying intertidal facies which can be helpful for paleo-sea level reconstructions (Neal, 2004; Van Heteren et al., 2000). However, the identification of the interface between intertidal and overlying aeolian sediments cannot be achieved easily on the basis of sedimentary structures and grain size analysis (Otvos, 2000). The paleo-shore line systems (i.e. shoreface-shelf) are significantly characterized into shallowing upward sequences bounded by marine flooding surfaces (Van Wagoner et al., 1990; Kamola and Van Wagoner, 1995).

2. STUDY AREA

2.1. Regional Setting

St. Joseph peninsula is situated on the panhandle of Florida west coast in the northeastern Gulf of Mexico at N29°50’ and W85°20’ (Fig. 1). It is located at the western edge of the Apalachicola Barrier Island Complex (ABIC; Fig. 2) which is along the North Florida Gulf of Mexico shoreline (Rink and López, 2010; Fig. 1). It extends to the SE-NW and is approximately 27km long with an average width of 305m (Fig. 2).

St. Joseph Peninsula (SJP) encloses a large bay on the north with separation from the Gulf of Mexico to the south with an approximate distance of 16km at its widest point (Figs. 2, 3). St. Joseph bay is unique, as it is the only sizable body of water along the northeastern Gulf of Mexico which is not of estuarine origin - i.e. not greatly affected by fresh water influx. St. Joseph bay was formed by the development of a cuspate spit (St. Joseph peninsula) and is being filled with Apalachicola river sand transported by longshore currents (Stewart et al., 1962). The windward side of the peninsula is bordered by beaches and dunes while the leeward side has tidal flats and marshes. Cape San Blas (CSB) is located on the south of the peninsula and at the southwestern edge of the westernmost portion of the ABIC (Figs. 2, 3). Similarly, CSB is characterized by the beach ridges and beach/dunes systems (Fig. 2,3).

2.2. Previous Work

Research on the evolution of the ABIC began in early 1960s when Tanner (1961) conducted research on the former Apalachicola River delta. Many studies relating to sedimentology, mineralogy, geomorphology, archaeology and sea-level history, have been conducted since then (e.g. Donoghue and Tanner, 1992; Rizk, 1991; Tanner et al., 1989). Many of these investigations have focused on the formation and evolution of ABIC (e.g. Stapor, 1975; Rizk, 1991; Otvos, 1992).

The first evolutionary model for the formation of the westernmost portion (St. Joseph Peninsula) of Apalachicola Barrier Island Complex (ABIC) was developed by Stapor (1973, 1975; Fig. 12). He proposed that Richardson's Hammock and the north side of Eagle Harbor were nuclei which were further connected by spit accretion through longshore currents (Fig. 12). Stapor (1975) reported a radiocarbon date (750 ^{14}C years B.P.) from a peat sample collected from the marshy area to the south of Richardson's Hammock. Rizk (1991) favored Stapor's model (1975) and examined the morphological characteristics (orientation, curvature and truncation) of individual beach ridges at St. Joseph peninsula, and divided these beach ridges into fourteen ridge sets. On the basis of these beach ridge sets, he determined the supra tidal progradation and evolution of the peninsula.

Stapor's model was further supported by more recent studies of Otvos (2005) where his four Thermo Luminescence (TL) ages (600 to 1700 years ago) agreed with the original model. Forrest (2003) found similar Optically Stimulated Luminescence (OSL) ages (n=4) that ranged from 8 to 1800 years BP and OSL ages from Rink and López (2010) supported that previous work.

López and Rink (2010) tested each section of the barrier (SJP) by dating sediment cores from fourteen locations. They employed the Single Aliquot Regenerative and Optically Stimulated Luminescence (OSL) dating techniques to determine the average sediment accumulation rate, progradation rates and ridge accretion rates. Generally, the oldest beach ridges are emplaced along the St. Joseph bay side whereas the youngest ridges are located close to the Gulf of Mexico shoreline. According to their proposed model, Richardson's Hammock which has the oldest ages and was the nucleus of the barrier with Rish Park and the area south of Eagle Harbor emerging at a later stage. They hypothesized that the existing barrier was formed by progradation and ultimate connection of these nuclei. They suggested that the CSB was formed, as a tombolo from the main land, at the last stages of this barrier system (i.e. less than about 1000 years ago). Their proposed model was in agreement with the Stapor's model - i.e. the Richardson's Hammock is the oldest land form and the St. Joseph Peninsula was formed after the formation of Richardson's Hammock.

3. METHODS

3.1. Field

Three vibra-cores were collected on the bay side of the peninsula (lee side) using a similar apparatus as Smith (1992) with sites selected to provide a shore perpendicular transect. Core 1 (N 29° 41’ 18.1284”, W 85° 19’ 42.3654”) was from the tidal flat area, Core 2 (N 29° 41’ 12.912”, W 85° 19’ 42.3654”) from the marsh and Core 3 (N 29° 41’ 08.916”, W 85° 19’ 42.2832”) from a swampy beach ridge swale (Fig. 3). Surface samples (n=18) were collected along a profiles across the CSB barrier (Fig.6). Elevations across the barrier were measured with a stadia and level and are referenced to Mean High Water (MHW) in the marsh (transition from high marsh to upland; Fig. 6). Four Ground Penetrating Radar GPR profiles were recorded using a sensors and software Noggin 250 instrument calibrated to 10m penetration at 250 MHz. Three shore perpendicular profiles were recorded on the gulf side of the barrier, with one of transects providing clear stratigraphic boundaries and is presented here (Fig.3).

3.2. Laboratory

Cores were split and peels were made using Avanti Hydro Sealant (AV-310) a hydrophilic polyurethane injection resin used to seal cracks in concrete. When it comes in contact with moisture, the resin expands and forms a flexible closed-cell polyurethane foam. The resin was mixed with equal parts water, stirred and then poured over exposed core surfaces and when cured (10 mins), peeled away. Differential penetration of the sealant into the sediment enhances sedimentary structures that cannot be normally seen (Fig. 4, 5). Peels were photographed and used for stratigraphic analysis. Cores were sampled at 1 cm resolution for particle-size, Loss on Ignition (LOI) and organic matter geochemistry.

3.2.1. Loss on Ignition (LOI)

LOI analysis followed the procedure used by Heiri et al., (2001). Samples were dried in an oven (Model - MDL 45EG) at 105°C for 12 hrs to remove moisture. Ignition of samples (at 550°C) was conducted for 2-3 hours in a Fisher Isotemp® 550 Series Muffle Furnace (model number 550-58) with the Organic Matter (OM) content calculated as weight percent (%). Results were plotted down core for analysis (Fig. 8).

3.2.2. Organic Matter Geochemistry

Samples for geochemical analysis were selected based on lithological units and to provide even distribution throughout the cores (Core 1, n=12; Core 2, n=15; Core 3, n=10; Fig. 8). The carbon isotopic composition ($^{13}C/^{12}C$ ratio, expressed as $\delta^{13}C$) and the organic carbon to total nitrogen ratio (C: N) samples were prepared and processed in Stable Isotope Research Laboratory (SIRL) at McMaster University. The samples were pretreated with 10% HCl for 12 hours to remove carbonates and rinsed with water. The samples were then dried at room temperature (≈ 25°C) for 24 hrs then crushed into finely powder form using mortar and pestle. Approximately 0.5 mg to 1.5 mg (depending on OM content) of sample was used for analysis. The geochemical analysis ($\delta^{13}C$ and C/N) were performed using a Costech Instruments Elemental Combustion System which was connected to a Thermo-Finnigan Delta Plus XP® mass spectrometer. The stable carbon isotope ratios were expressed in per mil (‰) relative to Vienna Pee Dee Belemnite (VPDB) against various standards (i.e. NBS-21, USGS-24, ANU-Sucrose, etc.). The uncertainty of standards for $\delta^{13}C$ was ± 0.1 ‰.

3.2.3. Textural Analysis

Particle-size analysis was conducted at 1 cm intervals on samples after combusted in the LOI procedure and thus contain only clastic components and shell material. Deionized water (1-2ml) was added to the samples and allowed to soak overnight. Samples were sand-sized, but were sieved to remove the >2000µm fraction (mostly large shells) before measurement. Particle-size measurement was conducted using a Beckman Coulter LS-230 laser particle-size analyzer using the Fraunhofer optical model to calculate Particle-Size Distributions (PSD) and statistical data. Approximately 1-2 ml of sodium hexametaphosphate solution (1-2%) was added as a dispersant, and samples homogenized by stirring as a moist paste before analysis (Donato et al., 2009).

Conventional textural parameters (mean, mode, standard deviation, skewness, and kurtosis) were calculated along with Particle Size Distribution (PSD) plots (Fig 8). PSDs for each core were plotted using Geosoft Oasis TM software (Beierle et al., 2002). Particle-size data was converted into phi log scale then gridded using the Triangular Irregular Network algorithm (TIN; Sambridge et al., 1995; Fig. 8). PSD surface plots allow visualization of minor particle-size variations in each core that may not be evident with normal statistics (i.e. mean, mode, standard deviation, skewness, and kurtosis; i.e. Inman, 1952, Folk and Ward, 1957). The particle size data, for each core was clustered using Ward's linkage and euclidean distance with the statistical program PAST^{TM}. This clustering produces a hierarchical dendrogram showing sample groupings that shared similar textural characteristics (Fig. 7). Hamilton (2007) provides a full description of the application of multi-variate statistics to particle-size distribution.

3.2.4. Radiocarbon Dating

Six samples including two from Core 1 and four from Core 2 were selected for radiocarbon dating (AMS; Fig. 8; Table 1). The samples were analyzed by Beta Analytic Inc. Miami, FL. Four of the samples were small twigs and the other two were of undifferentiated organic matter (see table 1).Samples were selected based on organic matter availability and also stratigraphic boundaries within the cores.

The samples were subjected to multiple pretreatments (i.e. acid/alkali/acid for twigs and acid washes for undifferentiated organic matter) prior to the analysis to eliminate contaminants. Conventional ages were then calibrated to years before present using INTCAL04 (Reimer et al., 2004; Talma et al., 1993; Table 1)

3.2.5. Paleontological Analysis

Whole and broken shell fragments (bivalves and gastropods) were collected from Core 1 and Core 2 and no shells were found in Core 3. Samples were washed and identified using Mikkelsen and Bieler (2007) and Warmke and Abbott (1961). For microfossil (foraminifera and thecamoebians) analysis, sub-samples of ≈ 4cc were sieved (38µm) and examined using an Olympus SZX 12 research stereomicroscope (model-ILLK100) at 60 to 90X. Thirty samples from regularly spaced intervals were examined (10 each core) but no foraminifera or thecamoebians were found in any of the samples.

4. RESULTS

4.1. Modern Facies

The CSB barrier has low-lying topography with maximum elevation at approximately 3 m above Mean-High Water (MHW; Fig. 6). The textural trends throughout the various sub-environments on the barrier (beach, dunes, beach ridges, swale swamps, marine marsh and intertidal sand flat) have relatively homogeneous textures. Sediment was predominantly sand-sized quartz. The beach or the windward side of the barrier has relatively smaller particle-sizes ≥ 3 □; v.f. sand) while the lee-side is slightly coarser at ≤ 3 □ (f. sand) but the difference is very small. Similarly, the mode and the Standard Deviations (SD) are very similar ranging from ≈ 1.5 to 2 □. The largest textural differences are with the swale swamps (mean = 8.6 □) and on the windward side, the beach flat and swash zone (mean = 1.0 to 1.5 □). The marine marsh also has some lower SDs at 2.5 □. Generally however, the sediment is a v. poorly sorted (≈ 3 □), fine sand with little variation across CSB.

4.1.1. Intertidal Sand flat

The tidal flat facies was characterized by light grey, m. sand sized (i.e. average mean = 1.9 □), very poorly sorted (i.e. average SD 3.1 □) sediment with abundant shell fragments (Fig. 6). Asymmetrical, straight crested tide dominated ripples were observed on the surface. However, no cross-bedding was evident in the short core taken at this facies environment. Bivalve shells (i.e. A*nomalocardia cuneimeris*) were dominant throughout this environment; most of the shells were fragmented (range: 1-10 cm) likely due to predation by birds and crabs. The proportion of angular shell fragments (i.e. 58%) was greater than that of the whole shells (i.e. 42%).

4.1.2. Marine Marsh

The marsh facies consists of light-dark grey; m. to f. sand sized (i.e. average mean = 2.0 □), very poorly sorted (i.e. average SD = 2.8 □) sediment with fibrous peat (LOI= 6 %; Fig. 6). *Spartina alterniflora/patens* occupied the low marsh whereas *Spartina patens / Juncus roemerianus* were dominant in the high marsh (Kemp et al., 2010). The marsh was approximately 80-100m wide and had cerithid gastropods and extensive burrows. The upland/high marsh facies was black, decomposed organic material and the vegetation transitioned to pine flatwoods and slash pine at higher elevation.

4.1.3. Upland - terrestrial

This facies is mainly consistent of beach ridges and dunes with variable amounts of vegetation ranging from sea-oats (i.e. *Uniola paniculata*) to forested areas (i.e. Pine flatwoods and slash pine etc.; Fig. 6). The beach ridge facies environment had an alternating sequence of ridges and swampy swales. The beach ridges and dunes were composed of light grey, very poorly sorted (i.e. average SD = 3.1 □), m. to f. sand sized (i.e. average mean = 1.9 □) sediment. The swampy swales had standing water as they were close to the water table and contained organic rich (50-60 %) sediment. The foredunes on the modern beach were variable in height but generally ranged from 0.5m to 1m. Roots casts and low-angle cross beds with burrows were evident in scarps in the dunes.

4.1.4. Beach

The beach contained m. to c. sand sized sediment (average mean = 1.2 □), very poorly sorted (average SD = 2.8 □; Fig. 6). Wrack lag lines on the beach had extensive accumulations of bored bivalve shell fragments. The swash zone sediments were slightly coarser (i.e. mean = 0.9 □), poorly sorted (SD = 1.62 □) and had abundant shell fragments. Gently dipping seaward cross laminations were found in short cores taken from this area.

4.2. Core Facies

The PSDs in all the cores were largely unimodal with a mode in most intervals between 1.7 and 2 ϕ and tended to be skewed with a fine tail into the silt range. Variations in the PSDs mostly occurred in those fine tails of the distribution with the dominant particle-size (mode approx. 2 ϕ) staying constant throughout the cores (Fig. 8).

Cluster analysis of the PSD data produced Litho-Facies (LF; intervals of similar PSDs) in the individual cores (Fig. 7). The similarity level of groupings was selected based on the lithologies visually logged with the sediment peels. This produced three clusters in Core 1, six clusters in Core 2 and six clusters in Core 3. These groupings largely followed the visual representation of facies within the cores confirming these assessments. In most instances, continuous sample intervals (e.g. samples from 72 - 116 cm in Core 2 Lithofacies 3) clustered together with very few groupings that contained mixtures of various odd samples (e.g. Core 2 Lithofacies 6; Fig 7b).

4.2.1. Swamp:

The Swamp Facies was found at the base of Core 1 and 2 (220-245 cm; 290-320 cm Fig. 8a, b) and at the top of Core 3 (0-60 cm; Fig. 8c). In Cores 1 and 2 it consists of OM sand with no discernible OM fragments (highly decomposed) whereas the unit at the top of Core 3 has extensive OM fragments (Sandy Woody Peat). The C:N ratios and $\delta^{13}C$ values in all three intervals are similar suggesting that they all formed in swamp. The C:N and $\delta^{13}C$ values (~12 to 30; -21 to -26 ‰) likely represent a mixture of C3 plants and freshwater DOC, POC and algae which are driving the C:N ratios to smaller values (Fig. 9). The Swamp Facies is part of the Strandplain Facies (discussed below) as it does not contain any shell material to suggest a marine origin and the ^{14}C dates are ~30 Ka years BP and > 44 Ka years BP and therefore formed during the last glacial when sea-level was much lower.

The textural characteristics of the Swamp Facies are very similar between the two units in Cores 1 and 2. In Core 1 (LF 3) and Core 2 (LF 6), the Swamp Facies clustered with parts of the Strandplain (Fig 7a, b; Fig 8a, b), but it clustered as a separate entity in Core 3 (LF 1 and 2, Fig.8c). The average mean particle size and mode are the same for facies in Cores 1 and 2 (1.8 □; 1.9 □) and the SD is slightly different at 2.8 and 2.9 □ respectively. The Swamp Facies in Core 3 also clusters separately from the rest of the core (LF1 and 2) but it is different in that it has textures dominantly in the silt range (mean = 6.0 □, mode = 5.3 □, SD = 6.5 □) and has higher LOI values (60%) than Cores 1 and 2 (≈ 2%). This is due to the better preservation of OM in the recent Swamp Facies in Core 3 vs the decomposed OM in the older intervals in Cores 1 and 2.

However, despite these differences in texture and LOI, the similarity in the OM geochemistry suggests that the basal layers of Cores 1 and 2 represent a similar environment as that to the modern swamp as found in Core 3 (Fig. 10)

4.2.2. Strandplain:

The designation of the Strandplain Facies is mostly inferred by the old radiocarbon dates obtained in Cores 1 and 2 (Core 1: 26 Ka years BP; Core 2: 12.8 ka years BP; Figs 8a, b) that indicate this facies cannot be of marine origin and must be terrestrial. Reinforcing this interpretation is the lack of shell material to suggest a marine environment which is seen in upper core intervals. The C:N and $\delta^{13}C$ are decreasing and increasing through this facies with the C:N ratios within the terrestrial plant range (C3 and C4) but the sediments also likely contain freshwater DOC, POC and algae (Fig. 9). The general increase in $\delta^{13}C$ values from the Swamp Facies could represent increasing C4 plant contributions and and/or decreasing influence of freshwater POC and algae indicative of sediment infilling the wetland. The "flaser-type" bedding seen at the base of the Strandplain Facies in the cores (Core 1: 140-220 cm; Core 2: 160-290 cm) is suggestive of alternating wet and dry conditions which maybe seasonal - this is also seen with the alternating LOI values (0.5 to 5%; Figs 8a, b). The OM rich units likely representing wet periods with ponding of water and the sand deposited during the dry months via wind or sheet wash during rainfall events. (Martin, 2000; Reineck and Wunderlich, 1968). The transition to more mottled OM sand maybe a reflection of the gradual infilling of the wetland and increased bioturbation (burrowing and roots).

The cluster analysis distinguished the Strandplain Facies in both Cores 1 (LF 2 and 3) and 2 (LF 4 to 6; Figs 7a, b). The PSDs tend to be uniform in both cores with some subtle cm scale changes occurring in the fine tails of the PSDs and there are regular alternations in the modal values between ~1.9 - 2.0 □ in both cores (Figs 8a, b). The transition from flaser-type bedding to more mottled OM sand is marked by decreasing OM contents in Core 1 (60-140 cm) but to a lesser degree in Core 2 (120- 155 cm).

The discordant radiocarbon dates 26 Ka vs 12.8 Ka at 115cm and 301 cm in Cores 1 and 2 (uncompacted) may indicate some reworking of the OM in the strandplain environment or the units accumulated at different times (Fig. 10). The reworking of OM is favored here as the cores are so close to each other (160 m) it does not seem plausible that the units accumulated 12 Ka apart. However, despite this discrepancy, there is nothing in the Strandplain Facies to indicate any marine influence and based on the radiocarbon dates they represent sediments deposited just prior to Holocene sea-level rise.

4.2.3. Intertidal Sand Flat:

The Intertidal Sand Flat Facies is marked by the appearance of shell material in the sediments (mainly *Anomalocardia cuneimeris* (bivalves), *Cerithium eburneum* (gastropods)) which begins at ≈ 60 cm in Core 1 and 120 cm in Core 2 (Figs 8a, b). LOI values drop to lower values (~ 0-2 % from ~ 2-3%) and mean particle size drops slightly in Core 1 but doesn't appreciably in Core 2 (i.e. vis a vis the Strandplain Facies). Cluster analysis results correspond to the appearance of the shell material in the cores (LF1 in Core 1; LF3 in Core 2; Figs 7a, b). The PSDs in this facies do not have a prominent fine tail as do the other facies (Strandplain). OM geochemistry in the Intertidal Sand Flat Facies is very similar to the underlying Strandplain Facies and may reflect reworking and mixing of older terrestrial OM with marine algae or POC, but the terrestrial OM seems to be dominating the values (Fig. 9).

The Storm Deposit Facies was distinguished as an individual lithofacies (LF2) in the cluster analysis but it is considered to be part of the Intertidal Sand Flat Facies (Core 2, ~ 40-70 cm). The Storm Deposit Facies does not contain abundant shell material, and is composed of slightly coarser sand (mean = 1.7 □, mode = 1.8□) likely coming from overwash as it does not contain any shell and is likely coming from terrestrial source but the LOI values are very low compared to the Marsh Facies (Figs 8a, b). C:N ratios increase and $\delta^{13}C$ values decrease compared to the underlying Intertidal Sand Flat Facies which likely reflects the intrusive roots which penetrated from the overlying Marsh Facies.

The ^{14}C date for the transition to marine sediments at 120 cm in Core 2 is ≈ 2.2 ka years BP with the transition in Core 1 inferred to be occurring at the same time or perhaps a little earlier as the cores are in close proximity to each other (160 m; Fig 10).

4.2.4. Marsh Facies:

The Marsh Facies is found in Core 2 and reflects the progradation of the lee-side of the CSB barrier (e.g. development of marsh over intertidal sand flat). This facies is characterized by fibrous sandy peat in Core 2 from the surface to 40 cm (Fig. 8b). The cluster analysis distinguished this facies from the Intertidal Sand Flat Facies (LF1) based on the PSDs which are not as prominently skewed with a fine-tail (Fig. 7b). The mean particle size is smaller (= 1.7 □) compared to the other facies but is very similar to the Storm Deposit Facies (Fig. 7b). The mode (1.8 □) and the SD are less variable in this facies (Fig. 8b). Geochemically, the Marsh Facies (Core 2) has slightly higher C:N ratios and the $\delta^{13}C$ values increase relative to the values observed in the Intertidal Sand Flat Facies (Fig. 9). Unfortunately, there is only one sample in this facies, but it indicates an increased marine influence as would be expected, with the values the sum of marine POC, algae, and terrestrial plant debris, sedges and marsh grasses (C3 - sedges e.g. *Scirpus olneyi* and C4 - marsh grasses e.g. *Spartina patens;* Arp et al., 1993).

The progradation of the marsh over the Intertidal Sand Flat Facies began around ≈ 340 years BP as found at the base of the fibrous peat in Core 2 at 40cm.

4.2.5. Aeolian Facies

The Aeolian Facies is found in Core 3 and is composed of medium sand (1-2□) with little variation in that size throughout the core except for the Swale Swamp Facies (Fig 8c). The cluster analysis separated the core into four lithofacies with LF 3 and 4 considered to be aeolian. The only variation between these two lithofacies was the extent of the roots traces which were more abundant in LF 3 and the presence of faint low-angle crossbeds as observed in the sediment peels. LF2 is a transitional unit between the aeolian sediments and the overlying swale swamp peat as the particle-size decreases (2 to 3 □) and the LOI values increase (0-30%) to LF1 values (Swale Swamp Facies). The PSDs in the Aeolian Dune Facies are different that the Strandplain Facies and the Marine Facies (Intertidal Sand flat, Storm Deposit and Marsh). The PSDs are skewed with a fine tail, but they also have multiple small peaks in the fine tail that are not displayed in any of the other facies (Fig 8c). Geochemically, the C:N ratios and the $\delta^{13}C$ values are all within the C3 terrestrial plant range which would be expected with grasses (e.g. sea-oats) and eventual forest cover and the formation of the swale swamp with rising sea-level and water table (Fig. 9).

4.3. Ground Penetrating Radar (GPR)

The GPR profiles show four distinct units within 3-4 m depth below the ground surface on windward beach of CSB (Figs. 3 and 11).

4.3.1. Unit A:

This is the deepest unit recorded in the profile, characterized by weak and numerous reflectors. This unit represents a shallow water table containing intrusive salt water which rapidly diminishes the propagating EM waves from GPR, thereby constraining the depth of penetration. Reflector surfaces and depositional geometry cannot be distinguished below this depth (i.e. > 3-4m; Fig. 11).

4.3.2. Unit B:

This unit is located approximately 1-3m below the ground surface. The unit contains distinct prograding clinoforms which downlap onto Unit A. The geometrical appearance of the clinoforms provide strong evidence of aggregation towards south (i.e. seaward). The aggregating clinoforms seem to be overlain by an erosional surface separating Unit B and Unit C (Fig. 11).

4.3.3. Unit C:

The base of this unit appears to be an erosional surface, truncating the underlying clinoforms of Unit B. Depositional structure within this unit contains a number of small, low angle reflectors which resemble beach dune deposits. These deposits are truncated by another erosional surface which acts as the top bounding surface of Unit C and separates it from the overlying Unit D. This top erosional surface contains multiple ridges and swales reflecting paleo-beach topography. The erosional nature and shallow depth (i.e. < 1m) might be a reflection of erosion due to high energy events (i.e. storms; Fig. 11).

4.3.4. Unit D:

This unit, separated by an erosional surface from the underlying unit C, is referred to the current modern beach deposits. Low-angle onlap and downlap terminations at the base suggest migrating dune field deposits towards the north (i.e. landward; Fig. 11).

5. DISCUSSION

5.1. Barrier Evolution

There is nothing in our data to suggest that there was a higher than present sea-level over the Holocene. Our stratigraphic data shows a slow rise in sea-level that flooded the older terrestrial surface at approximately 2.2 Ka. The elevation of this flooding surface in Core 2 (-1.4 m uncompacted depth; Fig. 10) is consistent with previous sea-level data collected in the Gulf Coast from submerged sites (Tornqvist et al., 2006; Wright et al., 2005; Milliken et al., 2008) which is at odds with the evidence for a mid-Holocene highstand coming from subaerial beach ridges (e.g. Stapor et al., 1991;

Tanner, 1992; Morton et al., 2000; Fig.14). The transition between the older Strandplain Facies (ca. 12 Ka) and the marine Intertidal Sand Flat Facies represents a ravinement surface that partially eroded the older sediments during the transgression (Fig. 13). Progradation with the Marsh Facies began by at least 0.3 Ka (Fig. 13).

The GPR and previous OSL dates largely confirms this model of barrier formation by providing data for the Gulf side of CSB and estimates of beach ridge formation (Rink and Lopez, 2010; Fig. 12). The GPR data shows the erosional truncation between Units B and D which likely corresponds to the ravinement surface dated to 2.2 Ka in the cores - the two are at approximately the same elevation (≈ -1.2 to 1.4 m) and likely occurred around the same time (Fig. 14). Unit B likely represents older sediments (>12 Ka Strandplain Facies) that are seen at the base of the cores.

Evidence on the formation of the barrier through the progradation of beach ridges is provided through OSL dates in Rink and Lopez (2010; Fig 12). The beach ridges started forming by at least 0.6 Ka and likely started sometime earlier as they did not date the most northerly ridges. Three other dates from CSB at 330 and 530 years BP show the subsequent progression of ridges to the present. This timing fits well with the radiocarbon ages and sequence in the intertidal cores where progradation of the Marsh Facies began at least by 0.3 Ka and likely earlier.

The data from the CSB indicates that at least this section, and likely others have an older core at their foundation and that the post-Holocene rise in sea-level sediments is a relatively thin veneer. Most of the ages obtained from the SJP and CSB are from relatively shallow (above sea-level) aeolian contexts (with possibly some shallow shoreface) and they are all relatively young (< 2 Ka). Several areas such as Richardson's Hammock and Rish Park have older ages (≈ 3 Ka) and are viewed as being the nucleus of the barrier (Rink and Lopez, 2010; Stapor 1973, 1975; Fig 12). However, in none of the previous research were dates found as old as those in this study (>12 Ka). They have not been reported as most of the previous cores probably did not penetrate deep enough to sample through the aeolian sediments, but it is likely that the whole barrier is underlain with these older sediments. The underlying core of the CSB is old (>12 Ka) but sea-level rise modified that surface after 2.2 Ka with the barrier forming sometime before 0.6 Ka with beach ridge accretion. Our results still reinforce the model proposed by Rink and Lopez (2010; modified from Stapor, 1973,1975); i.e. Richardson's Hammock is the nucleus (3.3 Ka) with Rish Park (2.2 Ka) and the area south of Eagle Harbor emerging later and CSB forming as a tombolo from the main land.

Older ages are not unfounded in previous studies in the area. Osterman et al., 2009 found ages ~ 40 to 54 Ka years BP in some of their cores from Apalachicola Bay just to the east of CSB. They interpreted the lowstand area to be a wooded vegetated coastal plain which fits our observations in our cores.

5.2. PSDs, Cluster Analysis and OM Geochemistry

The high resolution particle-size analysis generated a large data set of textural information for facies analysis. However, traditional particle-size statistics often oversimplifies the presentation of that data, missing important characteristics such as multi-modal distributions. PSD plots were developed to display this data in a format that would allow the full PSD to be seen (Beierle et al., 2002; van Hengstum et al., 2007). Multivariate analysis (Q-mode) allowed the large dataset of PSDs to be statistically compared and classified for facies analysis. Donato et al. (2009) found it worked effectively for classifying tsunami deposits in a coastal lagoon in Oman, but here we apply it to longer cores. In this study, cluster analysis divided the cores into distinct textural groupings that followed the other proxies providing greater confidence in facies designations.

Carbon isotopic ($\delta^{13}C$) and C:N ratios of coastal sediments were employed here as an alternate proxy for microfossils to categorizing the sub-tidal, inter-tidal and supra-tidal zones (Shennan et al., 2000). Often, the different sources of carbon input to the coastal and marine environments can be identified by using the isotopic variations of organic carbon (Megens et al., 2002; Lamb et al., 2006). However, the geochemical proxies were not overly effective here as there was some difficulty isolating the exact sources of OM. The geochemistry seemed to be effective in determining the OM rich environments such as the swamp and the marine marsh, but other facies were not as easily interpreted (Strandplain Facies). The variable inputs of OM in nearshore environments and reworking and mixing of older sediments provide a myriad of problems for the usage of OM geochemistry and as pointed out by Lamb et al., (2006) it is the relative change through the succession that is most important and that did prove effective in our cores.

6. Conclusions

The combined use of high resolution particle-size (PSDs), cluster analysis, isotope geochemistry and GPR profiles provided a comprehensive assessment of the stratigraphic development of the CSB in the absence of microfossil evidence. It is unclear why no foraminifera or thecamoebains were found in the cores in particular since they were found in the cores from Apalachicola Bay (Osterman et al., 2009).

The stratigraphic analysis found that the CSB was based on an older nucleus of strandplain deposits dating to >12 Ka that were subsequently flooded and modified through Holocene sea-level rise at ≈ 2.2 Ka. Actual barrier formation began sometime between 2.2.Ka and 0.6 Ka. However, Richardson's Hammock is emerged as a nucleus around 3 Ka which is the oldest beach ridge measured by Rink and Lopez (2010). Progradation of the barrier on the St Joseph Bay side began at least by 0.3 Ka and likely earlier. There is no evidence to indicate a higher than present sea-level in our core data and our data follows that of other studies using submerged samples (offshore).

LITERATURE CITED:-

ALEXANDER, C.S., 1969. Beach-ridges in Northeastern Tanzania: Geographical Review, v. LIX, pp.104-122.

ARP, B. G., DRAKE, W. T., POCKMAN, P. S., CURTIS and D. F., 1993. Whigham Interactions between C_3 and C_4 Salt Marsh Plant Species during Four Years of Exposure to Elevated Atmospheric CO2. Vegetation vol. 104/105, pp. 133-143.

BARBOZA, E.G., DILLENBURG, S.R., ROSA, M.L.C.C., TOMAZELLI, L.J., HESP, P.A., 2009. Ground-Penetrating Radar Profiles of Two Holocene Regressive Barriers in Southern Brazil. Journal of Coastal Research Special Issue 56, pp.579-583.

BEHRENS, E.W., 1966. Recent emerged beach in eastern Mexico: Science, 152, pp. 642–643.

BEIERLE, B.D., LAMOUREUX, S.F., COCKBURN, J.M.H., SPOONER, I., 2002. A New Method for Visualizing Sediment Particle Size Distribution. J. Paleolimnol. 27, pp.279–283.

BENNETT, M.R., CASSIDY, N.J., PILE, J., 2009. Internal Structure of a Barrier Beach as Revealed By Ground Penetrating Radar (GPR): Chesil Beach, UK. Geomorphology 104, pp.218-229.

BIGARELLA, J., 1965. Sand-ridge Structures from Parana Coastal Plain: Marine Geology, v.3, pp.269-278.

BLUM, M.D., MISNER, T.J., COLLINS, E.S., SCOTT, D.B., MORTON, R.A., and ASLAN, A., 2001. Middle Holocene Sea-Level Rise and Highstand at +2 m, Central Texas Coast: Journal of Sedimentary Research, v. 71, pp.581–588.

BLUM, M.D., CARTER, A.E., ZAYAC, T., and GOBLE, R., 2002. Middle Holocene Sea-Level and Evolution of the Gulf of Mexico Coast (USA): Journal of Coastal Research, Special Issue, v. 36, pp.65–80.

BLUM, M.D., SIVERS, A.E., ZAYAC, T., AND GOBLE, R.J., 2003. Middle Holocene Sea Level and Evolution of the Gulf of Mexico Coast: Transactions—Gulf Coast Association of Geological Societies, v. 53, pp.64–77.

CARTER, R.W.G., 1986. The Morphodynamics of Beach-ridge Formation; Magilligan, Northern Ireland: Marine Geology, v.73, pp.191-214.

COLEMAN, J.M., and SMITH, W.G., 1964. Late Recent rise of sea level: Geological Society of America Bulletin, v. 75, pp.833–840, DOI: 10.1130/0016–7606 (1964) 75 [833:LRROSL] 2.0.CO;2.

CURRAY, J.R., EMMEL, F.J., and CRAMPTONP, .J.S. Holocene History of a Strand Plain, Lagoonal Coast, Nayarit, Mexico. In, Costonaresa, A. and Phelgerf, B., eds., Lagunas Costeras, UN Simposio: Universidad National Autonoma de Mexico, 1967, 686p.

DAVIES, J.L., 1961. Tasmanian Beach-ridge System in Relation to Sea-Level Change: Proc. R. Soc. Tasmania, v.95, pp.35-40.

DAVIS JR., R.A., HINE, A.C., 1989. Quaternary Geology and Sedimentology of the Barrier Island and Marshy Coast, West central Florida, U.S.A. Field trip Guidebook T375, Int. Geol. Congress, AGU, Washington, DC.

DAVIS Jr., R.A. (Ed.), 1994. Geology of Holocene Barrier Island Systems. Springer, Berlin.

DE BEAUMONT, L.E., 1845. Septieme Lecon. In: Bertrand, P. (Ed.), LecVons de Ge¤ologie Pratique, pp. 221-252. Reprinted in: SCHWARTZ, M.L., Barrier Islands, Benchmark Papers in Geology. Dowden, Hutchinson and Ross, Stroudsburg, PA, pp. 5-43.

DEINES, P., 1980. The isotopic composition of reduced organic carbon. In: Fritz, P., Fontes, J.C. (Eds.), Handbook of Environmental Isotope Geochemistry. The Terrestrial Environment, vol. 1. A. Elsevier, Amsterdam, pp.329–406.

DONATO, S.V.; REINHARDT, E.G.; BOYCE, J.I.; PILARCZYK, J.E.; JUPP, B.P., 2009. Particle-size Distribution of Inferred Tsunami Deposit in Sur Lagoon, Sultanate of Oman. Marine Geology, 257, pp.54-64.

DONOGHUE, J.F., and TANNER, W.F., 1992. Quaternary terraces and shorelines of the panhandle Florida region, In Fletcher, C.H., and Wehmiller, J.F., eds., Quaternary Coasts of the United States: Marine and Lacustrine Systems: SEPM, Special Publication 48, pp.233–241.

DONOGHUE, J.F., and WHITE, N.M., 1995. Late Holocene sea-level change and delta migration, Apalachicola River region, northwest Florida, USA. J Coastal Res 11: pp.651–663.

DONOGHUE, J.F., STAPOR, F.W., and TANNER, W.F., 1998. Discussion of: Otvos, E.G., 1995. Multiple Pliocene-Quaternary marine highstands, northeast Gulf Coastal plain—Fallacies and facts: Journal of Coastal Research, v. 14, pp.669–674.

DONOGHUE, J.F. AND BALSILLIE, J.H., 2004. High resolution sea-level history for the Gulf of Mexico since the last glacial maximum. Florida Geological Survey. Report Investigation Number 103, pp.1-78.

EADIE, B. J., and JEFFREY, L. M., 1973. $\delta^{13}C$ analyses of oceanic particulate organic matter. Marine Chemistry I, pp.199-209.

FAIRBRIDGE, R. W., 1961, Eustatic change in sea-level: In: (L. H. Ahrens *et al.*, ed.), Physics and Chemistry of the Earth, v. 4, New York: Pergamon Press, pp.99-185.

FISHER, J.J., 1968. Barrier Island Formation: Discussion. Geol. Soc. Am. Bull. 79, pp.1421-1426.

FISHER, J.J., 1982. Barrier islands. In: Schwartz, M.L. (Ed.), The Encyclopedia of Beaches and Coastal Environments. Dowden, Hutchinson and Ross, Stroudsburg, PA, pp.124-134.

FOLK, R.L., WARD, W., 1957. Brazos river bar (Texas); A study in the significance of grain size parameters. J. Sediment. Petrol. 27, pp.3–26.

FOLK, R.L., 1966. A Review of Grain-Size Parameters. Sedimentology 6: pp.73–93.

FORREST, B.M., 2003. Applications of Luminescence Techniques to Coastal Studies, St. Joseph Peninsula, Gulf County, Florida. Unpublished M.Sc. Thesis, McMaster University, Hamilton, 209p.

FRIEDMAN, G. M., 1961. Distinction between Dune, Beach and River Sands from their Textural Characteristics. J. Sed. Pet. 31: pp.514–529.

FRY, B., SHERR, E.B., 1989. $\delta^{13}C$ Measurements as Indicators of Carbon Flow in Marine and Freshwater Ecosystems. In: RUNDEL, P.W., EHLERINGER, J.R., NAGY, K.A. (Eds.), Stable Isotopes in Ecological Research. Springer-Verlag, New York, pp.196–229.

GEHRELS, W.R. 1994. Determining relative sea-level change from salt-marsh foraminifera and plant zones on the coast of Maine, USA. Journal of Coastal Research 10, pp.990-1009.

GEHRELS, W.R., 2000. Using foraminiferal transfer functions to produce high-resolution sea-level records from salt-marsh deposits Maine, USA. The Holocene 10 (3), pp.367–376.

GEHRELS, W.R., ROE, H.M., CHARMAN, D.J., 2001. Foraminifera, Testate Amoebae and Diatoms as sea-level indicators in UK saltmarshes: a quantitative multiproxy approach. J. Quat. Sci. 16 (3), pp.201–220.

GIBBS, A.E., DAVIS, R.A., 1991. Development and Stratigraphy Three Rooker Bar: A Recently Emergent Barrier Island, Pinellas County, Florida. 12th Annu. Res. Conf. SEPM Gulf Coast Section, pp.1-7.

GILBERT, G.K., 1885. The Topographic Features of Lake Shores. 122 R.A. DAVIS Jr. et al. /Marine Geology 200 (2003) 103-123 US Geological Survey 5th Annual Report, pp.

87-88. Reprinted in: SCHWARTZ, M.L. (ED.), Barrier Islands, Benchmark Papers in Geology. DOWDEN, HUTCHINSON and ROSS, Stroudsburg, PA, pp.45-46.

GOMEZ-ORTIZ, D., PEREIRA, M., MARTIN-CRESPO, T., RIAL, F.I., NOVO, A., LORENZO, H., VIDAL, J.R., 2009. Joint Use of GPR and ERI to Image the Subsoil Structure in a Sandy Coastal Environment. Journal of Coastal Research Special Issue 56, pp.956-960.

GONZALEZ, S., INNES, J., HUDDART, D., DAVEY, P., PLATER, A.J., 2000. Holocene Coastal Change in the North of the Isle of Man: Stratigraphy, Palaeoenvironment and Archaeological Evidence. In: Pye, K., ALLEN, J.R.L. (EDs.), Coastal and Estuarine Environments: Sedimentology, Geomorphology and Geoarchaeology. Geological Society, London, Special Publications, vol. 175, pp. 343–363.

HAMILTON, L.J., 2007. Clustering of cumulative grain size distribution curves for shallow marine samples with software program CLARA. Aust. J. Earth Sci. 54, pp.503–519.

HAMPSON, G.J., 2000. Discontinuity Surfaces, Clinoforms, and Facies Architecture in a Wave-Dominated, Shoreface–Shelf Parasequence: Journal of Sedimentary Research, v. 70, pp.325–340.

HORTON, B.P., 1999. The distribution of contemporary intertidal foraminifera at Cowpens Marsh, Tees Estuary, UK: implication for studies of Holocene sea-level changes. Palaeogeogr. Palaeoclimatol. Palaeoecol. 149, pp.127–149.

HORTON, B.P. and EDWARDS, R.J. 2006. Quantifying Holocene sea-level change using intertidal foraminifera: Lessons for the British Isles. Cushman Foundation for Foraminiferal Research, Special Publication 40, 97p.

HEIRI, O., LOTTER A.F. and LEMCKE G. 2001. Loss on Ignition as a Method for Estimating Organic and Carbonate Content in Sediments: Reproducibility and Comparability of Results. J. Paleolim. 25: pp.101–110.

INMAN, D.L., 1952. Measures for Describing the Size Distribution of Sediments. Journal of Sedimentary Petrology 22: pp.125–145.

JOHNSON, D.W., 1919. Shore Processes and Shoreline Development, Wiley, New York.

KAMOLA, D.L., and VAN WAGONER, J.C., 1995. Stratigraphy and facies architecture of parasequences with examples from the Spring Canyon Member, Blackhawk Formation, Utah, in Van Wagoner, J.C., and Bertram, G.T., eds., Sequence Stratigraphy of Foreland Basin Deposits: American Association of Petroleum Geologists, Memoir 64, pp.27–54.

KEMP, A.C., HORTON, B.P. and CULVER, S.J. 2010. Stable carbon isotopes as potential sea level indicators in salt marshes, North Carolina, USA: Implications for sea-level research laboratory.pp.1-38.

KIDSON, C., 1982. Sea Level Changes in the Holocene. Quaternary Science Review, v. 1, pp. 121 - 151.

LAMB, A.L., WILSON, G.P. and LENG, M.J., 2006. A Review of Coastal Palaeoclimate and Relative Sea-Level Reconstructions using δ13C and C/N Ratios in Organic Material. Earth-Science Reviews 75, pp.29-57.

LAMB, A.L., VANE, C.H., WILSON, G.P., REES, J.G. and MOSS-HAYES, V.L. 2007. Assessing δ13C and C/N ratios from organic material archived in cores as Holocene sea level and palaeoenvironmental indicators in the Humber Estuary, UK. Marine Geology 244, pp.109-128.

LEATHERMAN, S.P., 1982. Barrier island handbook: Maryland, University of Maryland, 109p.

LOCKER, S.D., DOYLE, L.D., HINE, A.C., BLAKE, N.J., 1998. Complex carbonate and clastic stratigraphy on the inner shelf o¡ west-central Florida. Bull. Am. Assoc. Petrol. Geol. 74, 707p.

LOCKER, S.D., HINE, A.C., DAVIS, R.A., BROOKS, G.R., GUY, K.K., 2003. West-central Florida coastal studies project. U.S. Geol. Survey Open File Reports Nos. 99-505 to 99-513.

LÓPEZ, G.I., 2007. The Late Quaternary Evolution of the Apalachicola Barrier Island Complex, North-East Gulf of Mexico, as Determined from Optical Dating. Ph.D. Dissertation, School of Geography and Earth Sciences, McMaster University, Hamilton, Canada, 264p.

MARTIN, A.J. (2000). "Flaser and wavy bedding in ephemeral streams: a modern and an ancient example". Sedimentary Geology 136 (1-2): pp.1–5.

MASON, O.K., 1991. A Geoarchaeological Methodology for Studying Prograding Coastal Sequences: Beach-Ridge Geomorphology in Kotzebue Sound, Alaska, in: Paleoshorelines and Prehistory, Johnson, L.C., and Strightm, (eds), CRC Press, Boca Raton, Florida, pp.54-64.

MCFARLAN, E., Jr., 1961, Radiocarbon Dating of Late Quaternary Deposits, South Louisiana: Geological Society of America Bulletin, v. 72, p.129–158.

MCKEE, K.L., PATRICK, W.H., 1988. The Relationship of Smooth Cordgrass (Spartina alterniflora) to Tidal Datum: A Review. Estuaries 11, pp143– 151.

MCGEE, W.J., 1890. Encroachment of the Sea. In: Metcalf, L.S. (ED.), The Forum, vol. 9, pp. 437-449. Reprinted in: Schwartz, M.L., Barrier Islands, Benchmark Papers in Geology. Dowden, Hutchinson and Ross, Stroudsburg, PA, pp.49-61.

MEGENS, L., VAN DER PLICHT, J., DE LEEUW, J.W., SMEDES, F., 2002. Stable Carbon and Radiocarbon Isotope Compositions of Particle Size Fractions to Determine Origins of Sedimentary Organic Matter in an Estuary. Organic Geochemistry 33, pp.945–952.

MIDDELBURG, J.J., NIEUWENHUIZE, J., 1998. Carbon and nitrogen stable isotopes in suspended matter and sediments from the Scheldt Estuary. Marine Chemistry 60, pp.217– 225.

MEYERS, P.A., 1994. Preservation of elemental and isotopic source identification of sedimentary organic matter. Chemical Geology 114, pp.289–302.

MIKKELSEN, P.M., BIELER, R., 2007. Seashells of southern Florida: living marine mollusks of the Florida Keys and adjacent regions: bivalves. Princeton University Press, Princeton, NJ.

MILLIKEN, K.T., ANDERSON, J.B., and RODRIGUEZ, A.B., 2008. A New Composite Holocene Sea-Level Curve for the Northern Gulf of Mexico, in ANDERSON J.B., and MOIOLA, R.J. and WEISER, D.L., 1968. Textural Parameters: An Evaluation. Journal of Sedimentary Petrology v. 38, pp.57–69.

MORTON, R.A., PAINE, J.G., and BLUM, M.D., 2000. Responses of Stable Bay-Margin and Barrier-Island Systems to Holocene Sea-Level Highstands, Western Gulf of México: Journal of Sedimentary Research, v. 70, pp.478–490. Downloaded from GSA bulletin.gsapubs.org on January 6, 2010.

NEAL, A., PONTEE, N.I., PYE. K., RICHARDS, J., 2002. Internal Structure of Mixed-Sand-and-Gravel Beach Deposits Revealed Using Ground-Penetrating Radar. Sedimentology v.49, pp.789-804.

NEAL. A., 2004. Ground-Penetrating Radar and its Use in Sedimentology: Principles, Problems and Progress. Earth-Science Reviews 66, pp.261-330.

NEILSON, 2009. Long term post-orogenic evolution of N Atlantic conjugate margins constrained by on and off shore data. Volume V – pp.211-215.

NISSENBAUM, A. and KAPLAN, I. R., 1972. Chemical and isotopic evidence for the in situ origin of marine humic substances; Li and Wlogy and Oceanography 17, pp.560-582.

OTVOS, E.G., 1970. Development and Migration of Barrier Islands, Northern Gulf of Mexico. Geol. Soc. Am. Bull. 81, pp.241-246.

OTVOS, E.G., 1979. Barrier Island Evolution and History of Migration, North Central Gulf Coast. In: Leatherman, S.P. (ED.), Barrier Islands from the Gulf of St. Lawrence to the Gulf of Mexico, pp.291-319.

OTVOS, E.G., 1981. Barrier Island Formation Through Nearshore Aggradation-Stratigraphic and ¢eld Evidence. Mar. Geol. 43, pp.195-243.

OTVOS, E.G., 1984. Alternative Interpretations of Barrier Island Evolution, Apalachicola Coast, Northwest Florida. Litoralia 1, pp.9-21.

OTVOS, E.G., 1992. Apalachicola Coast Quaternary Evolution, NE Gulf of Mexico. In: CLETCHER, C.H., WEHMILLER, J.F. (EDs.), Quaternary Coasts of the United States: Marine and Lacustrine Systems. Society of Economic Paleontologists and Mineralogists, Tulsa, pp.221–232.

OTVOS, E.G., 1999. Sediment and geomorphic criteria for reconstructing sea-level positions, Multiple Pliocene-Quaternary marine highstands on the northeastern Gulf of Mexico coastal plain? : Reply to a discussion: Journal of Coastal Research, v. 15, pp.1181–1187.

OTVOS, E.G., 2000. Beach ridges—Definitions and Significance: Geomorphology, v. 32, pp.83–108, DOI: 10.1016/S0169-555X (99)00075-6.

OTVOS, E.G., 2001. Assumed Holocene highstands, Gulf of Mexico: basic issues of sedimentary and landform criteria—discussion: Journal of Sedimentary Research, v. 71, pp.645–647.

OTVOS, E.G., 2004. Holocene Gulf levels: recognition issues and an updated sea-level curve: Journal of Coastal Research, v. 20, pp.680–699.

OTVOS, E.G., 2005. Coastal barriers, Gulf of Mexico: Holocene evolution and chronology. Journal of Coastal Research, Special Issue 42, 141–163.

PATTERSON, R.T., et al., 2005. Relative utility of foraminifera, diatoms and macrophytes as high resolution indicators of paleo-sea level. Quat. Sci. Rev. 24, pp.2002–2014.

PENLAND, S., and RAMSEY, K.E., 1990. Relative Sea-Level Rise in Louisiana and the Gulf of Mexico: 1908–1988: Journal of Coastal Research, v. 6, pp.323–342.

PEKALA, J.M., 1996. Stratigraphy and Geologic History of Anna Maria Island, Manatee County, Florida. Unpublished Master's Thesis, Univ. of South Florida.

PENLAND, S., MCBRIDE, R.A., WILLIAMS, S.J., BOYD, R., and SUTER, J.R., 1991. Effects of sea level rise on the Mississippi River delta plain, in Kraus, N.C., Gingerich, K.J., and Kriebel, D.L., eds., Coastal sediments '91: New York, American Society of Civil Engineers, pp.1248–1264.

PETERS, K.E., SWEENEY, R.E., KAPLAN, I.R., 1978. Correlation of carbon and nitrogen stable isotopes in sedimentary organic matter. Limnology and Oceanography 23, pp.598– 604.

PIRAZZOLI, P.A., 1991. World Atlas of Holocene sea-level changes: Amsterdam, Elsevier, 300p.

PRAHL, F.G., BENNETT, J.T., CARPENTER, R., 1980. The early diagenesis of aliphatic hydrocarbons and organic matter in sedimentary particulates from Dabob Bay, Washington. Geochimica et Cosmochimica Acta 44, pp.1967–1976.

PSUTY, N.P., 1967. The Geomorphology of Beach-ridges in Tabasco, Mexico: Baton Rouge, LA, Louisiana State University Press, 51p.

PYE, K., and BLOTT, S.J., 2004. Particle Size Analysis of Sediments, Soils and related Particulate materials for Forensic Purposes using Laser Granulometry. Forensic Sci. Int., 144, pp.19–27.

RANDAZZO, A.F., and JONES, D. S. (Eds), 1997. The geology of Florida: Gainesville, University Press of Florida, 327p.

REDMAN, J.B., 1852. On the Alluvial Formations, and the local Changes of the South Coast of England: Minutes of Proceedings Institution of Civil Engineers, v. XI, pp.162-223.

REIMER, P.J., BAILLIE, M.G.L., BARD, E., BAYLISS, A., BECK, J.W., BERTRAND, C.J.H., BLACKWELL, P.G., BUCK, C.E., BURR, G.S., CUTLER, K.B., DAMON, P.E., EDWARDS, R.L., FAIRBANKS, R.G., FRIEDRICH, M., GUILDERSON1, T.P., HOGG, A.G., HUGHEN, K.A., KROMER, B., MCCORMAC, G., MANNING, S., RAMSEY, C.B., REIMER, R.W., REMMELE, S., SOUTHON, J.R., STUIVER, M., TALAMO, S., TAYLOR, F.W., VAN DER PLICHT, J., WEYHENMEYER, C.E., 2004. InTOCal04 terrestrial radiocarbon age calibration, 0-26 cal kyr BP. Radiocarbon 46 (3), pp.1029-1058.

REINECK, H.E.; WUNDERLICH, F. (1968). "Classification And Origin Of Flaser And Lenticular Bedding". Sedimentology 11 (1-2): pp.99–104.

REINHARDT, E.G., NAIRN, R., LOPEZ, G., 2010. Recovery estimates for the Río Cruces after the May 1960 Chilean earthquake, Marine Geology v. 269, pp.18–33 (JIF: 1.975)

RINK, W.J., and LÓPEZ, G.I., 2010. OSL-Based Lateral Progradation and Aeolian Sediment Accumulation Rates for the Apalachicola Barrier Island Complex, North Gulf of Mexico, Florida. Geomorphology, 123. pp. 330-342.

RIZK, F.F., 1991. The Late Holocene Development of St. Joseph Spit. Florida State University, Tallahassee. RODRIGUEZ, A.B., eds., Response of Upper Gulf Coast Estuaries to Holocene Climate Change and Sea Level Change: The Geological Society of America Special Paper 443, pp.1–11.

RODRIGUEZ, A.B., ANDERSON, J.B., SIRINGAN, F.P., and TAVIANI, M., 2004. Holocene evolution of the east Texas coast and inner continental shelf: along-strike variability in coastal retreat rates: Journal of Sedimentary Research, v. 74, pp.405–421.

RODRIGUEZ, A.B., and MEYER, C.T., 2006. Sea-level variation during the Holocene deduced from the morphological and stratigraphic evolution of Morgan Peninsula, Alabama, USA: Journal of Sedimentary Research, v. 76, pp.257–269.

SAMBRIDGE, M., BRAUN, J., MCQUEEN, H., 1995. Geophysical parameterization and interpolation of irregular data using natural neighbors. Geophys. J. Int. 122, pp.837–857.

SCHLEE, J.S., UCHUPI, E., and TRUMBALL, J.V.A., 1964. Statistical Parameters of Cape Cod Beach and Aeolian sands. U.S. Geol. Surv. Prof. Pap. 501-D: Dll8--D122.

SCHNABLE, J. E., and GOODELL, H. G., 1968, Pleistocene-Recent stratigraphy, evolution, and development of the Apalachicola coast, Florida: Geological Society of America Special Paper No. 112, 72p.

SCHWARTZ, M.L., 1971. The multiple causality of barrier islands. J. Geol. 79, 91-94.

SCOTT, D.K. and LECKIE, M.R.M., 1990. Foraminiferal zonation of Great Sippewissett Salt Marsh (Fahnouth, Massachusetts), J. Foraminiferal Res., 20: pp.248-266.

SCOTT D.B. and MEDIOLI, F.S., 1978. Vertical zonations of marsh foraminifera as accurate indicators of former sea levels. Nature, 272: pp.528-531.

SCOTT, D.B. and MEDIOLI, F.S., 1980. Quantitative studies of marsh foraminiferal distributions in Nova Scotia: implications for sea level studies. Cushman Found. Foraminiferal Res. Spec. Publ., 17, 58p.

SCOTT, D. B., MEDIOLI, F. S., and SCHAFER, C. T. (2001). Monitoring in coastal environments using foraminifera and thecamoebian indicators, Cambridge Univ Pr.

SHENNAN, I. 1986. Flandrian sea-level changes in the Fenland II: Tendencies of sea-level movement, altitudinal changes and local and regional factors. Journal of Quaternary Science 1, pp.156-178.

SHENNAN, I. 1992. Late Quaternary sea-level changes and crustal movements in eastern England and eastern Scotland: an assessment of models of coastal evolution. Quaternary International 15, pp.161-173.

SHENNAN, I., TOOLEY, M.J., GREEN, F., INNES, J., KENNINGTON, K., LLOYD, J., RUTHERFORD, M., 1999. Sea Level, Climate Change and Coastal Evolution in Morar, Northwest Scotland. Geologie en Mijnbouw 77, pp.247–262.

SHENNAN, I., LAMBECK, K., HORTON, B., INNES, J., LLOYD, J., MCARTHUR, J., PURCELL, T., RUTHERFORD, M., 2000. Late Devonian and Holocene Records of Relative Sea-level Changes in Northwest Scotland and their Implications for Glacio-hydro-isostatic Modelling. Quaternary Science Reviews 19, pp.1103– 1135.

SHEPARD, F.P. and YOUNG, R., 1961. Distinguishing between Beach and Dune Sands. J. Sediment. Petrol., 31: pp.196-214.

SHEPARD, F.P., 1964. Sea-level changes in the past 6000 years: Possible archaeological significance: Science, v. 143, pp.574–576, DOI: 10.1126/ science.143.3606.574.

SMITH, D.G. 1992. Vibracoring: recent innovations. J. Paleolim. 7: pp.137–143.

SOLOHUB, J.T. and KLOVAN, J.E., 1970. Evaluation of Grain-size Parameters in Lacustrine Environments. Journal of Sedimentary Petrology, 40(1), pp.81–101.

STAPOR, F.W., 1973. Coastal Sand Budgets and Holocene Beach Ridge Plain Development —Northwest Florida. Florida State University, Tallahassee. 221p.

STAPOR, F.W., 1975. Holocene Beach Ridge Plain Development, Northwest Florida. Zeitschrift für Geomorphologie 22, pp.116–144.

STAPOR, F. W., and TANNER, W. F., 1977. Late Holocene mean sea-level data from St. Vincent Island, and the shape of the late Holocene sea-level curve: In: (W. F. Tanner, ed.), Coastal Sedimentology, Tallahassee, FL: Department of Geology, Florida State University, pp. 35-68.

STAPOR, F.W., JR., MATHEWS, T.D., and LINDFORS-KEARNS, F.E., 1991. Barrier-island Progradation and Holocene sea-level history in southwest Florida: Journal of Coastal Research, v. 7, pp.815–838.

STAPOR, F. W., and STONE, G. W., 2004. A new depositional model for the buried 4000 BP New Orleans Barrier: implications for sea-level fluctuations and onshore transport from a nearshore shelf source: Marine Geology, v. 204, pp.215-234.

STEVENSON, J. C., WARD, L. G., and KEARNEY, M. S., 1986. Vertical Accretion in Marshes with Varying Rates of Sea- level Rise, in Estuarine Variability, edited by D. A. WOLFE, pp.241–259, Academic, Orlando, Fla.

STEWART, R. A., AND D. S. GORSLINE, 1962. Recent sedimentary history of St. Joseph Bay: Sedimentology, v. 1, pp.256---286.

TANNER, W.F., 1961. Offshore Shoals in area of Energy Deficit. Journal of Sedimentary Petrology 31, pp.87–95.

TALMA A.S., VOGEL J.C., 1993. A Simplified Approach to Calibrating ^{14}C Dates: Quaternary Dating Research Unit: Radiocarbon, v. 35, No.2, pp.317-322.

TANNER, W.F., 1970. Growth rates of Venezuelan Beach-ridges: Sedimentary Geology, v.6, pp.215-220.

TANNER, W.F., DEMIRPOLAT, S., STAPOR, F.W., and ALVAREZ, L., 1989. The ‘‘Gulf of Mexico’’ late Holocene sea level curve: Gulf Coast Association of Geological Societies, Transactions, v. 39, pp.553–562.

TANNER, W.F., 1992. Late Holocene sea-level changes from grain-size data: evidence from the Gulf of Mexico: The Holocene, v. 2, pp.249–254.

TANNER, W.F., 1995. Origin of beach ridges and swales: Marine Geology, v. 129, pp.149–161.

THOMAS, M.A., and ANDERSON, J.B., 1994. Sea-level controls on the facies architecture of the Trinity/Sabine incised valley system, Texas continental shelf, in Dalrymple, R.W., Boyd, R., and Zaitlin, B.A., eds., Incised-valley systems: Origin and sedimentary sequences: SEPM (Society for Sedimentary Geology) Special Publication 51, pp.63–82.

TÖRNQVIST, T.E., GONZALEZ, J.L., NEWSOM, L.A., VAN DER BORG, K., DE JONG, A.F.M., and KURNIK, C.W., 2004. Deciphering Holocene sea-level history on the U.S. Gulf Coast: A high resolution record from the Mississippi Delta: Geological Society of America Bulletin, v. 116, p. 1026–1039, DOI: 10.1130/B2525478.1.

TÖRNQVIST, T.E., BICK, S.J., VAN DER BORG, K., and DE JONG, A.F.M., 2006. How Stable is the Mississippi Delta? Geology, v. 34, pp.697–700, DOI: 10.1130/ G22624.1.

TYSON, R.V., 1995. Sedimentary Organic Matter: Organic Facies and Palynofacies. Chapman and Hall, London.

VAIL, P. R., 1987. Part 1: Seismic Stratigraphy Interpretation Procedure. In: Bally, A.W. (ED.), Atlas of Seismic Stratigraphy: AAPG Studies in Geology 1(27), pp.1-10.

VAN DE PLASSCHE, O. (Ed.) 1986. Sea-level research: a Manual for the Collection and Evaluation of Data. Geo Books.

VAN HENGSTUM, P.J., REINHARDT, E.G., BOYCE, J.I., CLARK, C., 2007. Changing sedimentation patterns due to historical land-use change in Frenchman's Bay, Pickering, Canada: evidence from high- resolution textural analysis. J. Paleolimnol. 37, pp.603–618.

VAN HETEREN, S., HUNTLEY, D.J., VAN DE PLASSCHE, O., and LUBBERTS, R.K., 2000. Optical dating of dune sand for the study of sea-level change: Geology, v. 28, pp.411–413.

VAN WAGONER, J.C., MITCHUM, R.M., JR., CAMPION, K.M., and RAHMAN-IAN, V.D., 1990. Siliciclastic Sequence Stratigraphy in Well Logs, Core, and Outcrops: Concepts for High-Resolution Correlation of Time and Facies: American Association of Petroleum Geologists, Methods in Exploration Series 7, 55p.

WADA, E., MINAGAWA, M., MIZUTANI, H., TSUJI, T., IMAIZUMI, R., KARASAWA, K., 1987. Biogeochemical studies on the transport of organic matter along the Otsuchi River watershed, Japan. Estuarine, Coastal and Shelf Science 25, pp.321–336.

WARMKE, G.L., ABBOTT, R.T., 1961. Caribbean Seashells; A guide to the marine mollusks of Puerto Rico and other West Indian Islands, Bermuda and the Lower Florida Keys. Livingston Pub. Co, Narberth, Pennsylvania.

WILSON, G.P., LAMB, A.L., LENG, M.J., GONZALEZ, S. and HUDDART, D., 2005. Variability of organic $\delta^{13}C$ and C/N in the Mersey Estuary, U.K. and its implications for sea-level reconstruction studies. Estuarine, Coastal and Shelf Science 64, pp.685-698.

WILSON, G.P., LAMB, A.L., LENG, M.J., GONZALEZ, S., HUDDART, D., 2005b. $\delta^{13}C$ and C/N as Potential Coastal Paleo-environmental Indicators in the Mersey Estuary, UK. Quaternary Science Reviews 24, pp.2015–2029.

WRIGHT, L.D., 1970. The Influence of Sediment Availability on the Patterns of Beach-ridge Development in the Vicinity of the Shoal haven River Delta, N.S.W: The Australian Geographer, v.11, pp.336-348.

WRIGHT, E.E., HINE, A.C., GOODBRED, S.L., and LOCKER, S.D., 2005. The Effect of Sea-level and Climate Change on the Development of a mixed Silici-clastic Carbonate, Deltaic Coastline: Suwannee River, Florida, USA: Journal of Sedimentary Research, v. 75, pp.621–635, DOI: 10.2110/jsr.2005.051

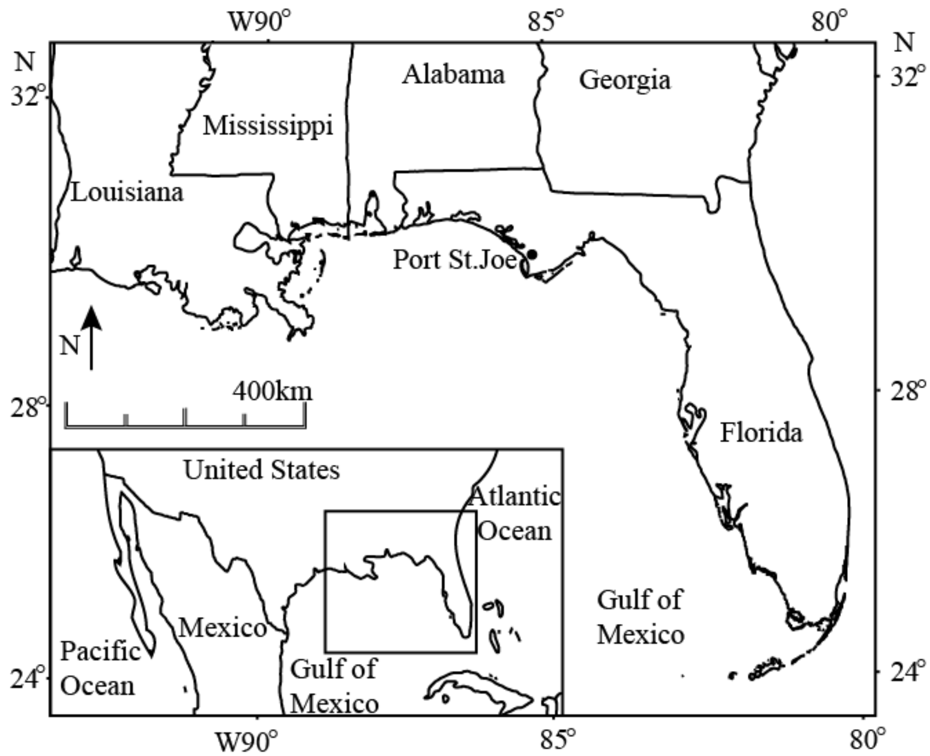

Figure 1. Location of Florida relative to the Atlantic Ocean and Gulf of Mexico.

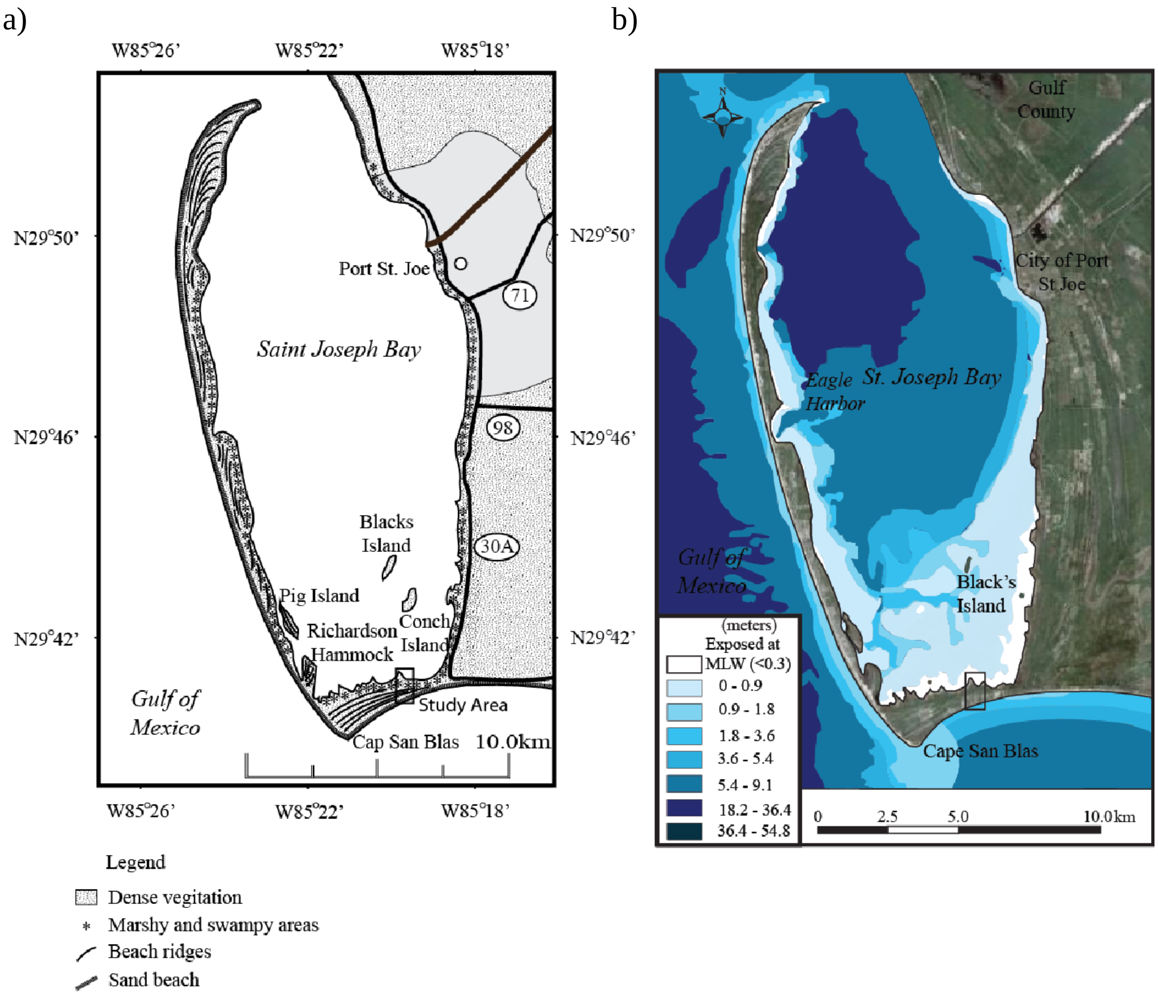

Figure 2. a) Location of study area relative to Port St. Joe and Gulf of Mexico. b) Digital Elevation Model (DEM) of the study area.

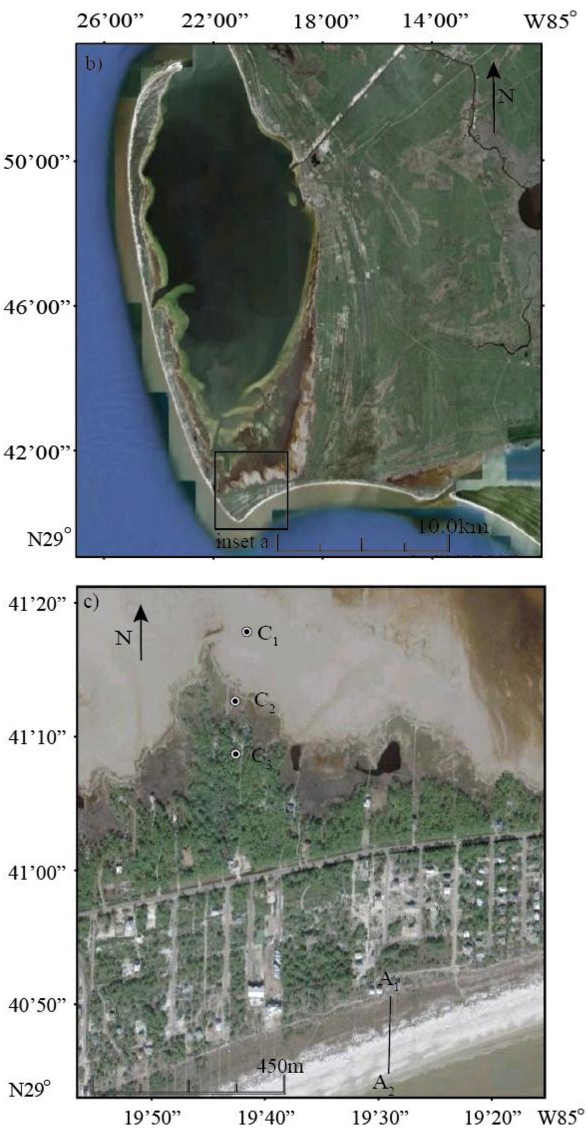

Figure 3. Aerial view of the study area showing locations of cores (C_{1-3}) taken in St. Joseph Bay and GPR profile (A_1-A_2) in Cape San Blas, Florida.

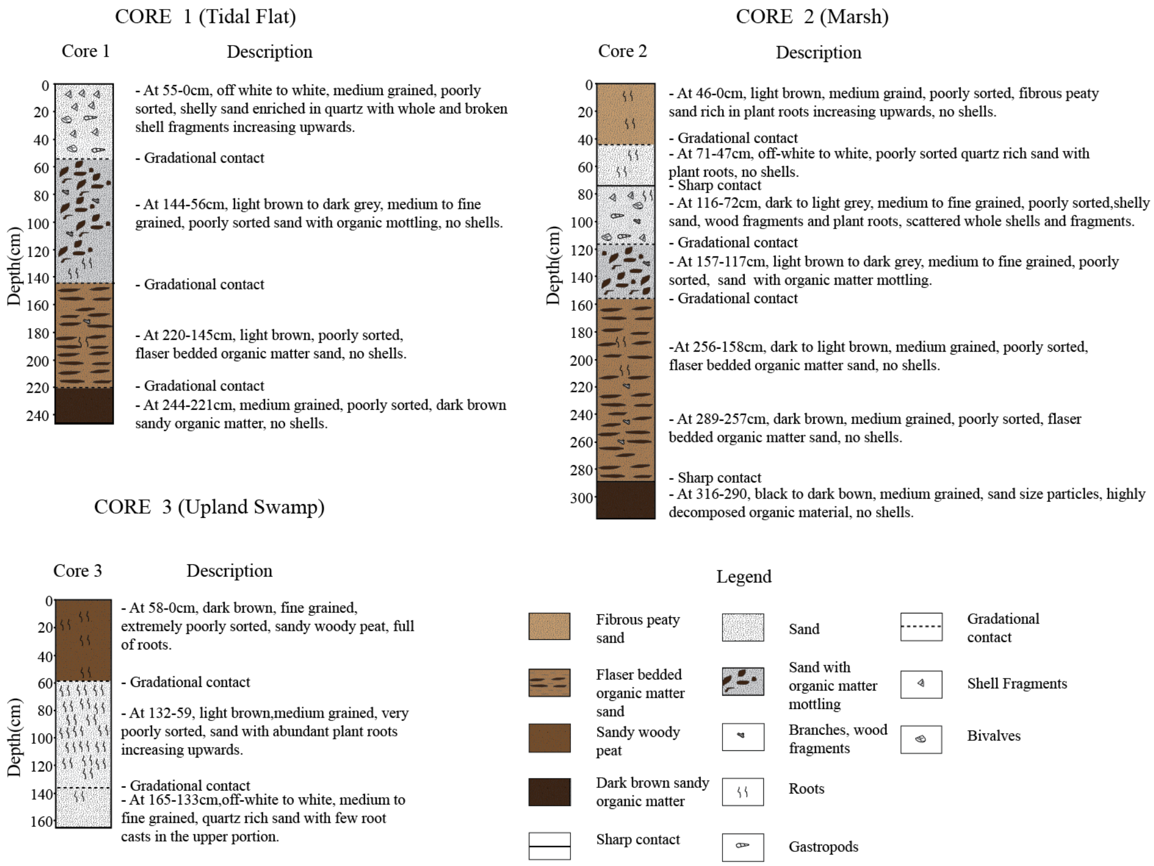

Figure 4. The lithological descriptions of cores.

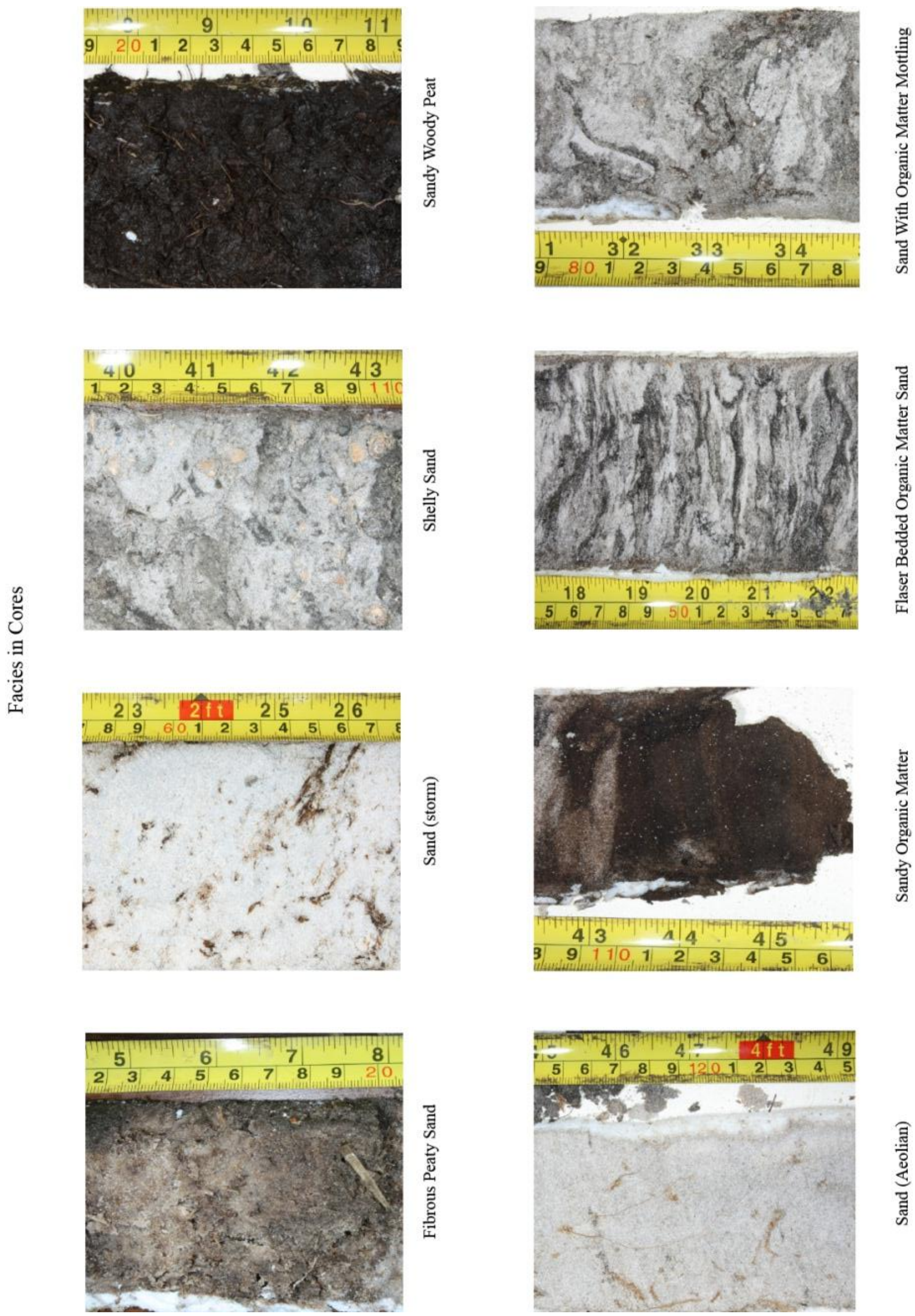

Figure 5. Representative photography of lithological units (sediment peels).

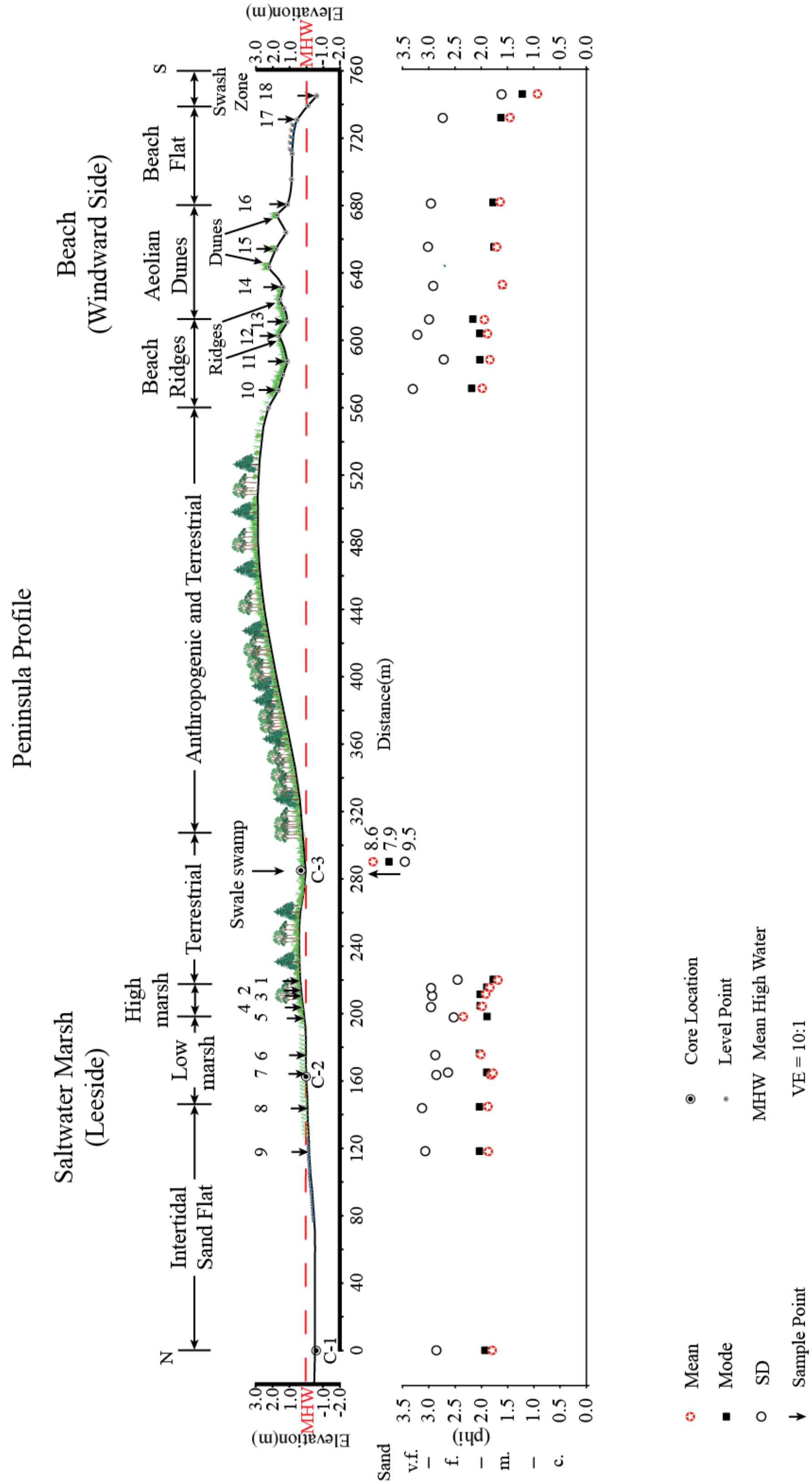

Figure 6. Cross-sectional profile of Saint Joseph Peninsula based on leveling points, showing the locations of cores and the surface samples.

a)

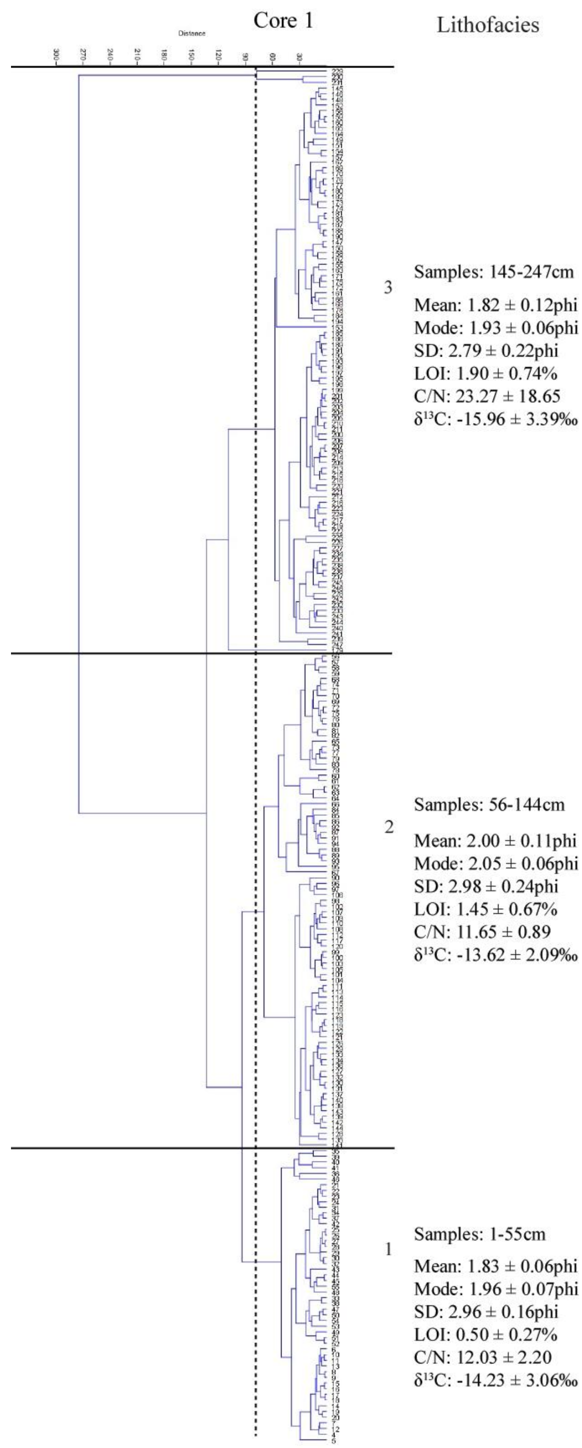

Figure 7a. Q-mode cluster results showing mean values and sample interval for each lithofacies in Core 1.

b)

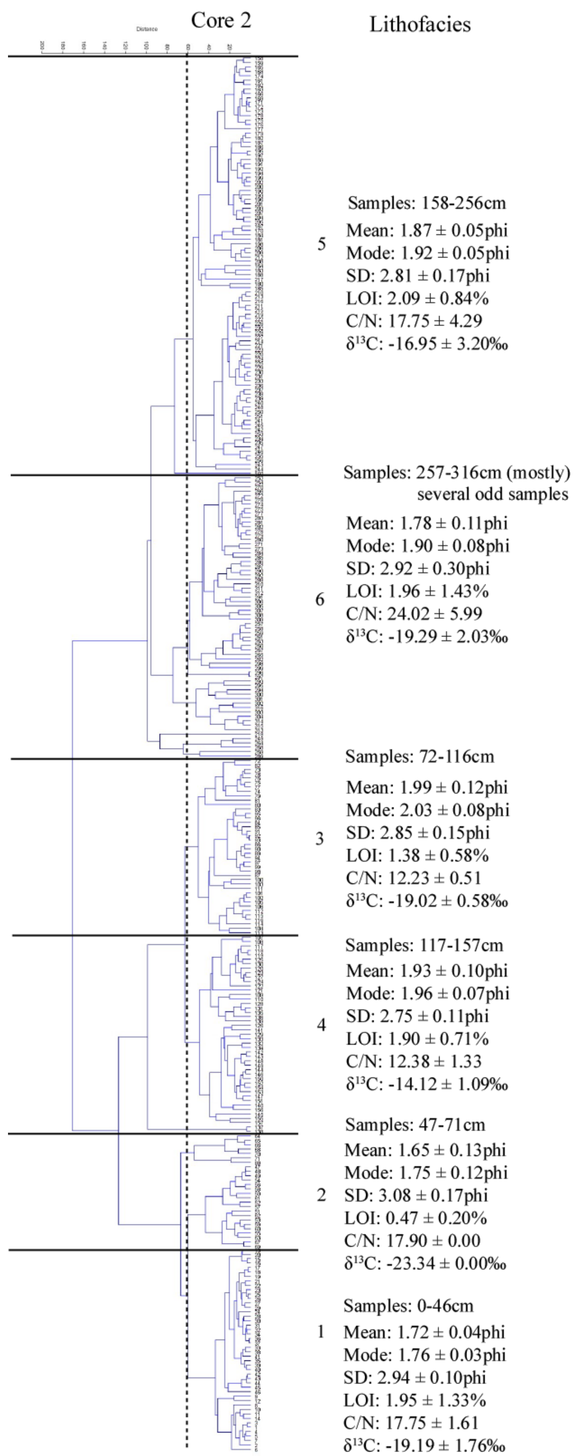

Figure 7b. Q-mode cluster results showing mean values and sample interval for each lithofacies in Core 2.

c)

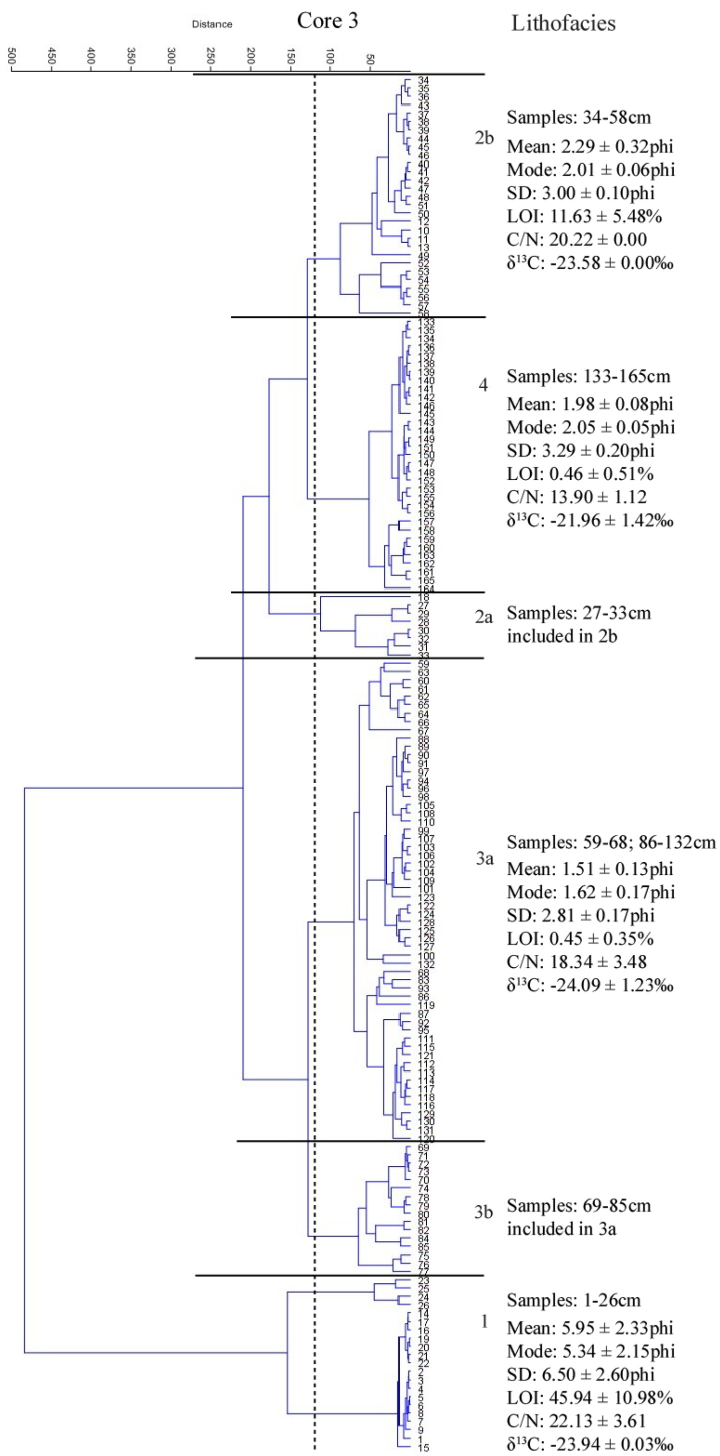

Figure 7c. Q-mode cluster results showing mean values and sample interval for each lithofacies in Core 3.

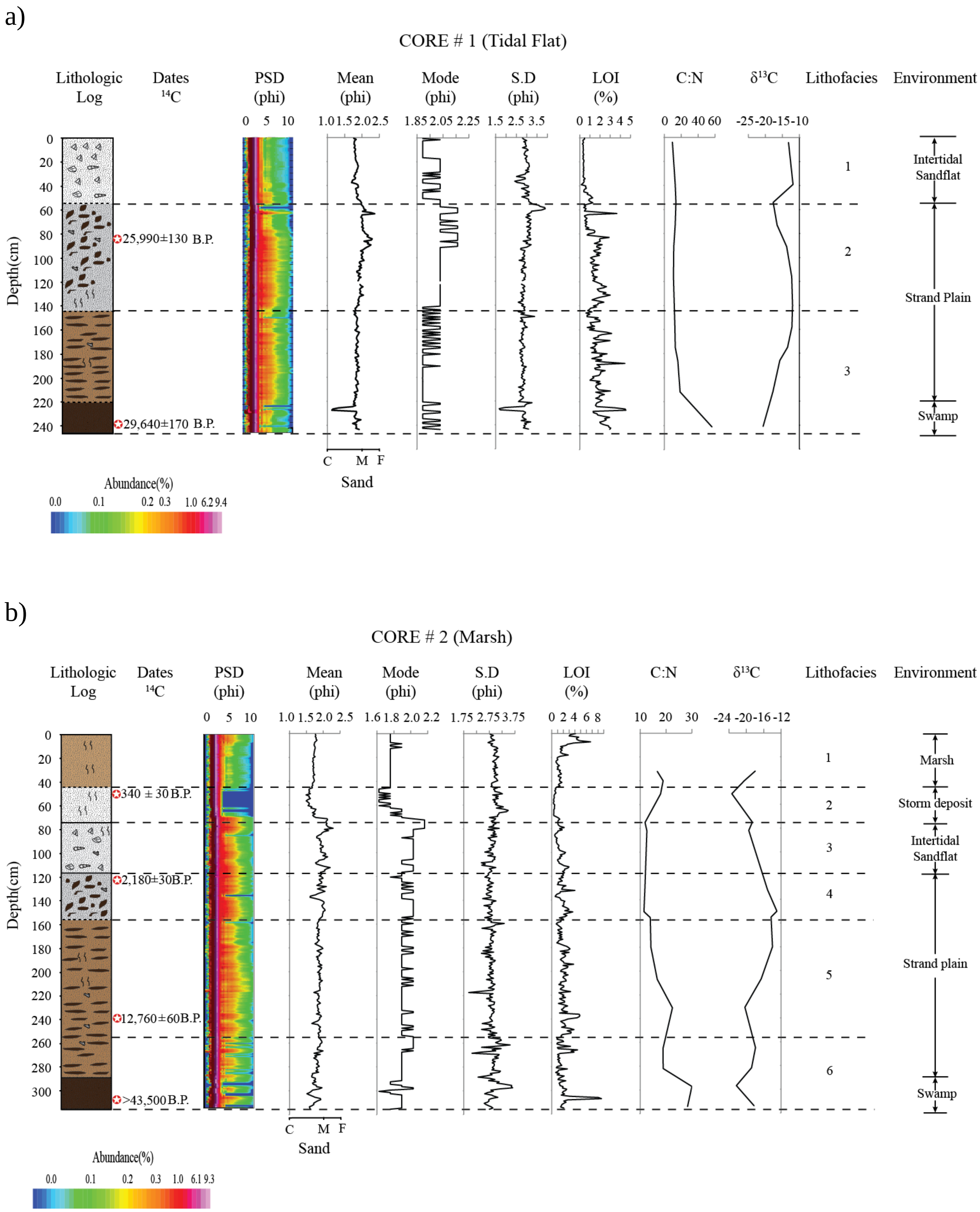

Figure 8 (a, b). Lithologic logs, particle size distributions (PSD), conventional statistics (mean, mode and standard deviation), LOI, C:N, δ13C, cluster analysis defined lithofacies for compacted cores (1,2). See Fig.5 for facies photographs.

c)

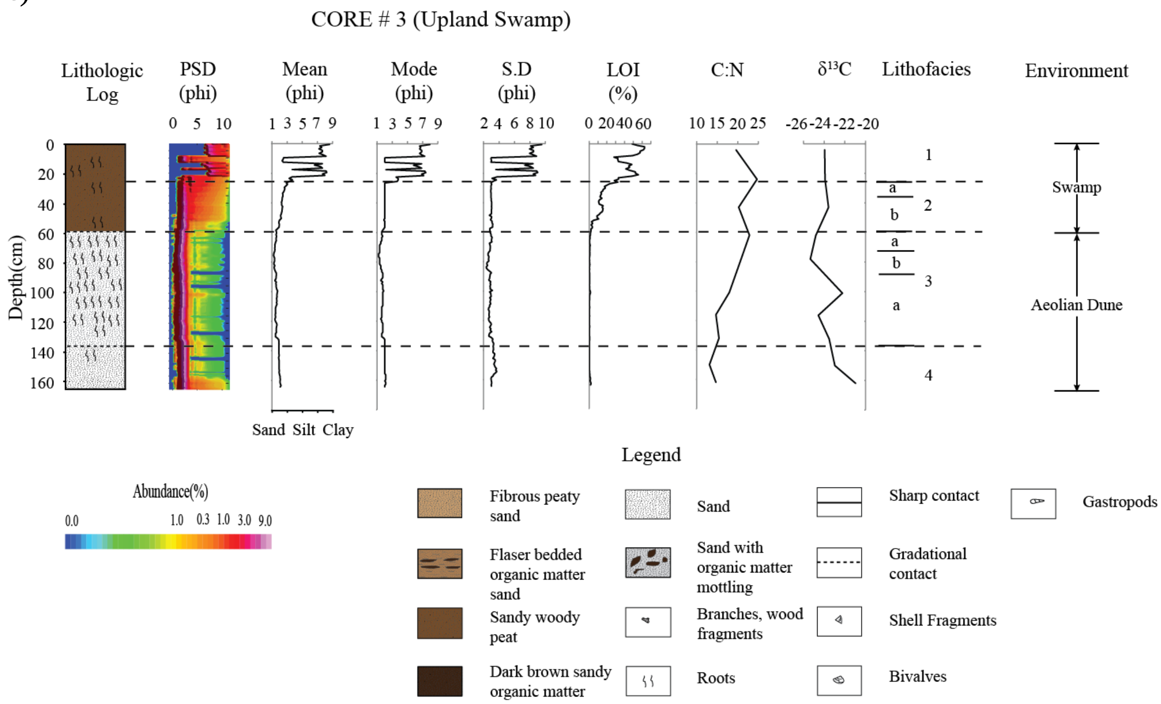

Figure 8c. Lithologic logs, particle size distributions (PSD), conventional statistics (mean, mode and standard deviation), LOI, C:N, $\delta^{13}C$, cluster analysis defined lithofacies for compacted core 3. See Fig.5 for facies photographs.

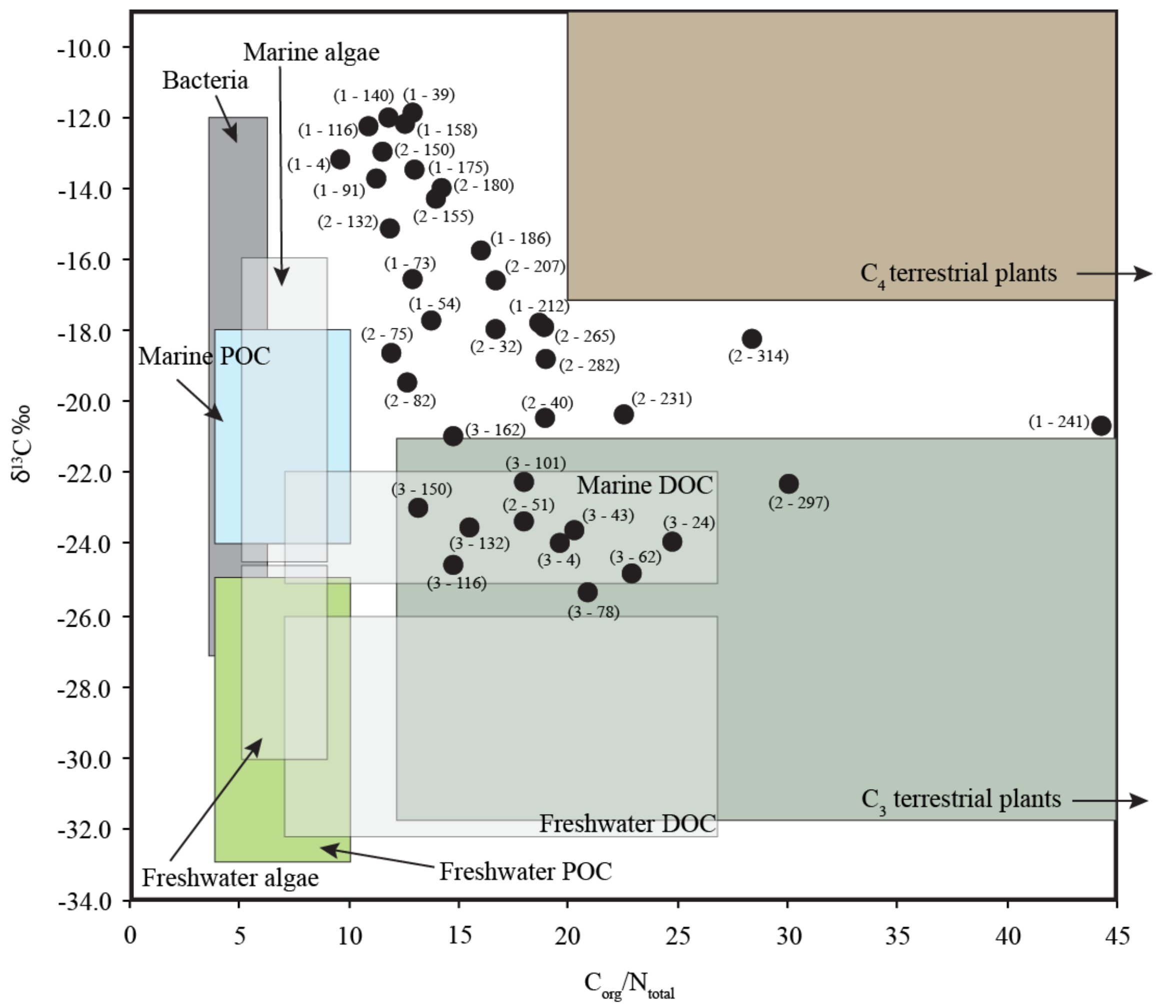

Figure 9. Plot showing distribution of $\delta^{13}C$ and C/N ratios and the typical $\delta^{13}C$ and C/N ranges for organic inputs to coastal environments after Lamb et al., 2006 (numbers in brackets refer to core and sample depth). POC – Particulate Organic Carbon; DOC – Dissolved Organic Carbon.

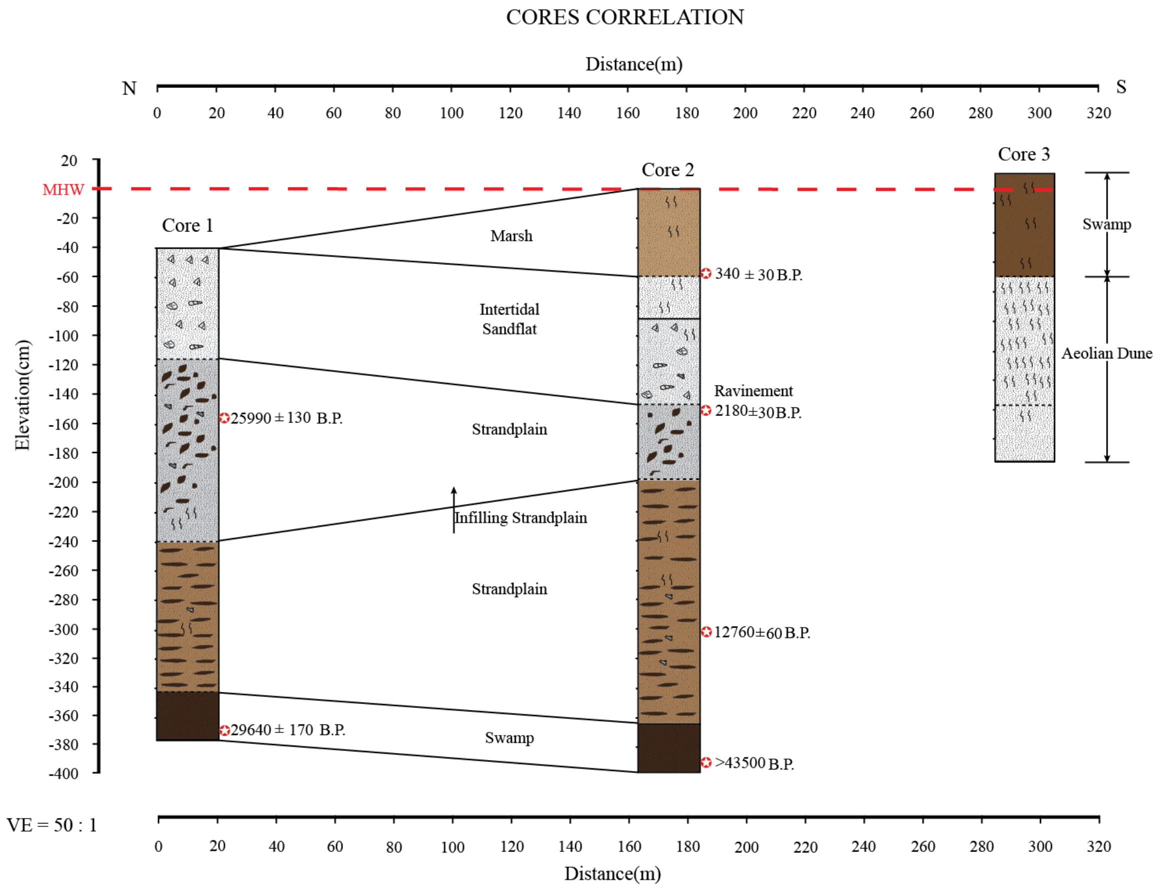

Figure 10. Correlation of the three vibra-cores (uncompacted) based on lithofacies and organic matter geochemical analysis. Core locations are shown in Figures 3 and 6.

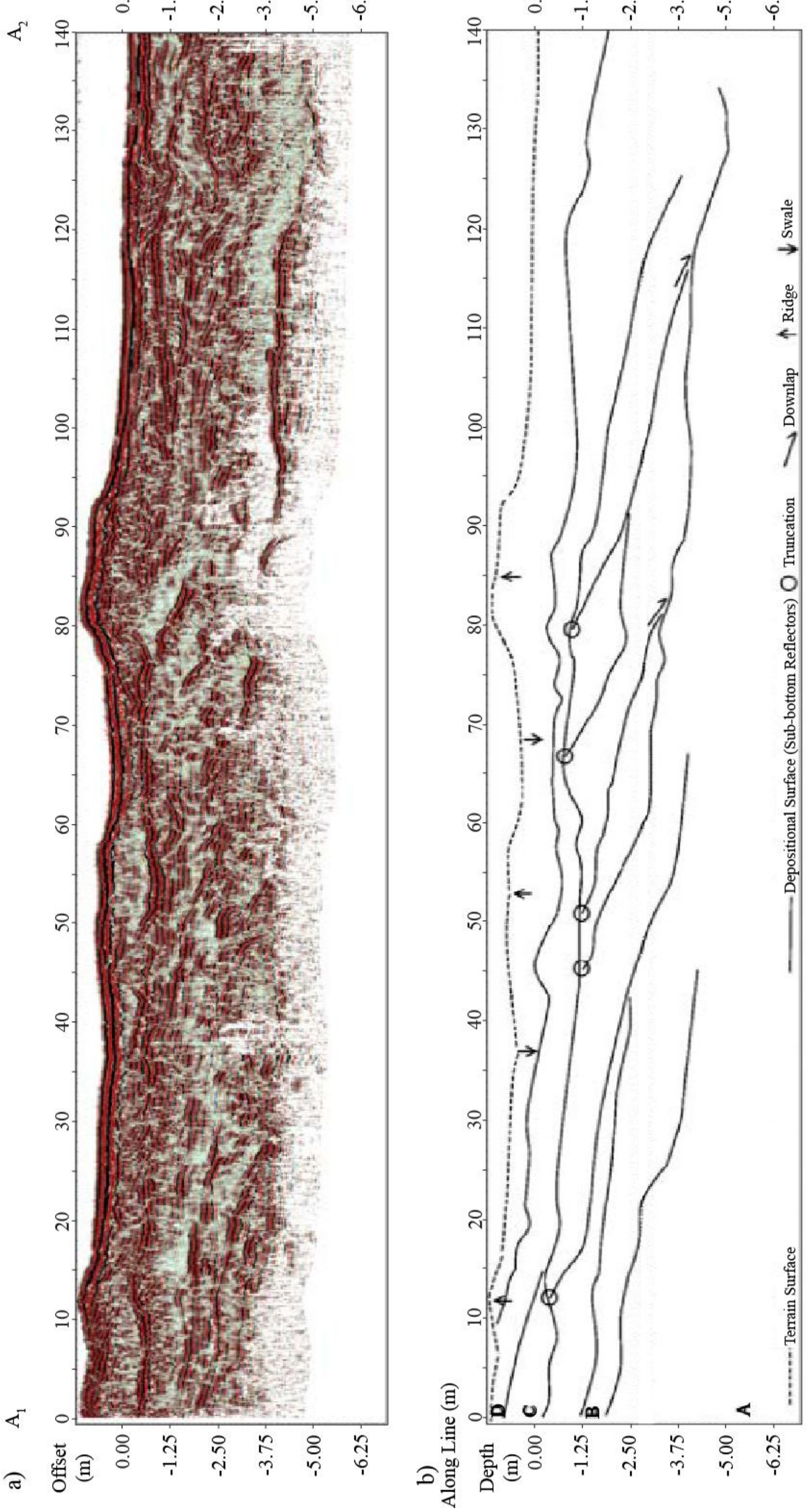

Figure 11. A) Processed Cape San Blas GPR profile 2 at 250 MHz. B) Interpreted sub-surface geometry and depositional surfaces seen on radargram A (Fig. 3).

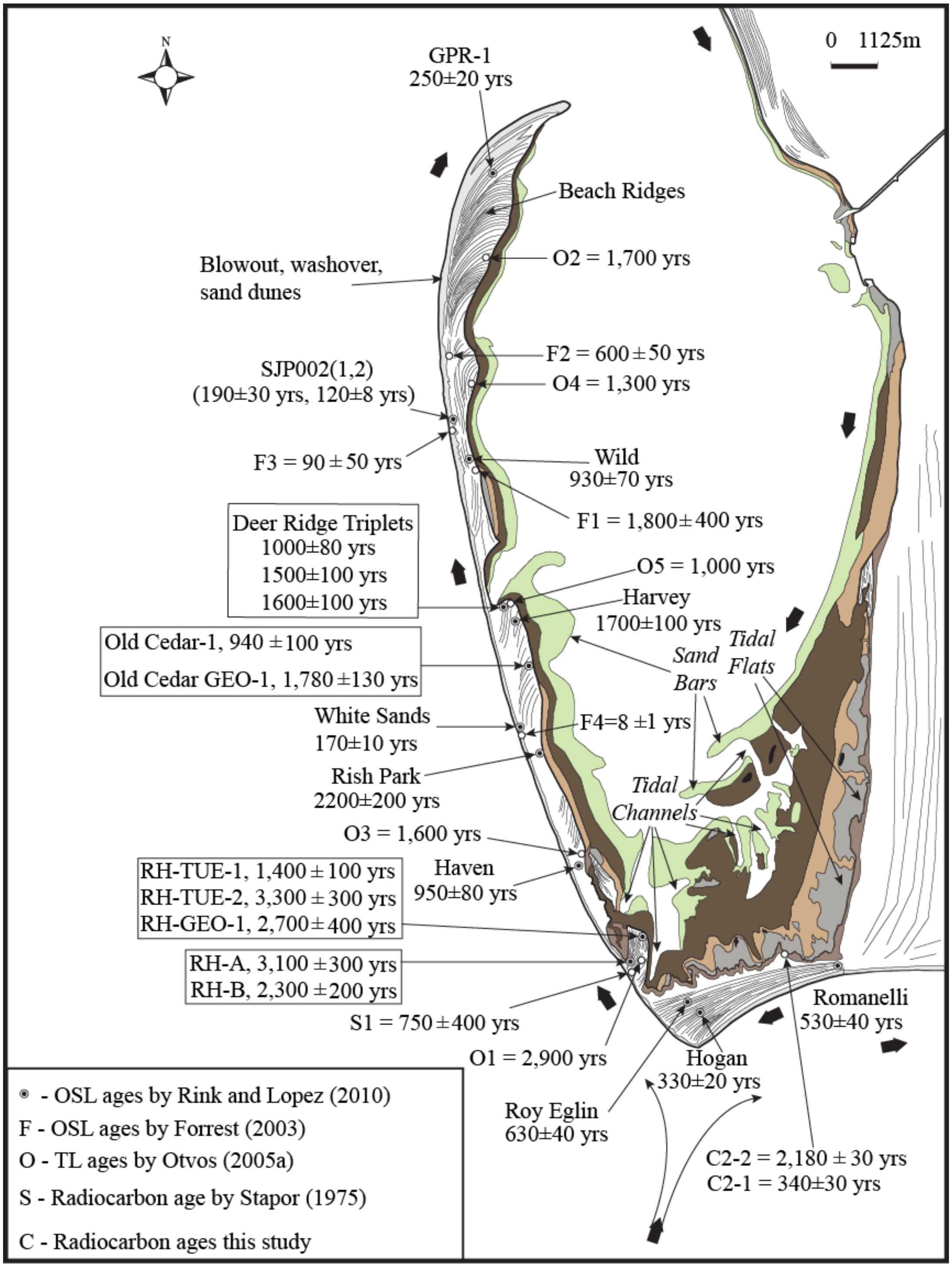

Figure 12. Map of Saint Joseph peninsula (modified after Rink and Lopez, 2010) showing compiled OSL dates: (Stapor 2007, Forrest 2003, Otvos 2005a, Rink and Lopez 2010) and (^{14}C) from this study.

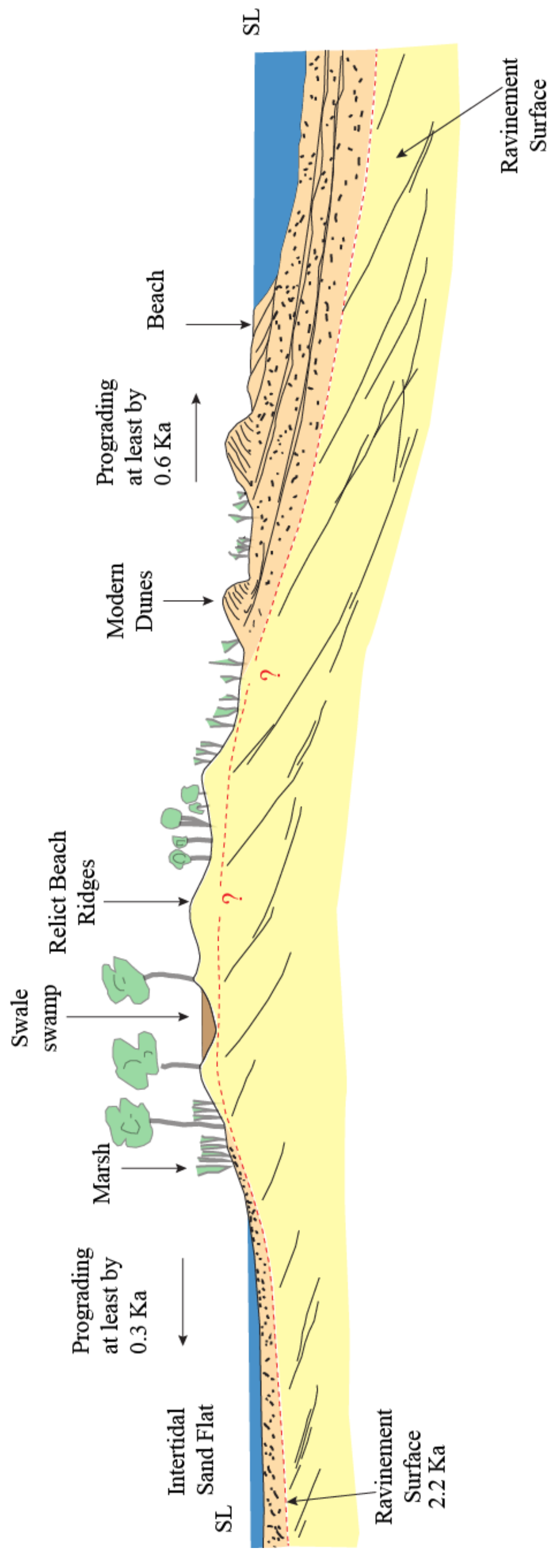

Figure 13. Interpreted cross-section through barrier showing ravinement surface.

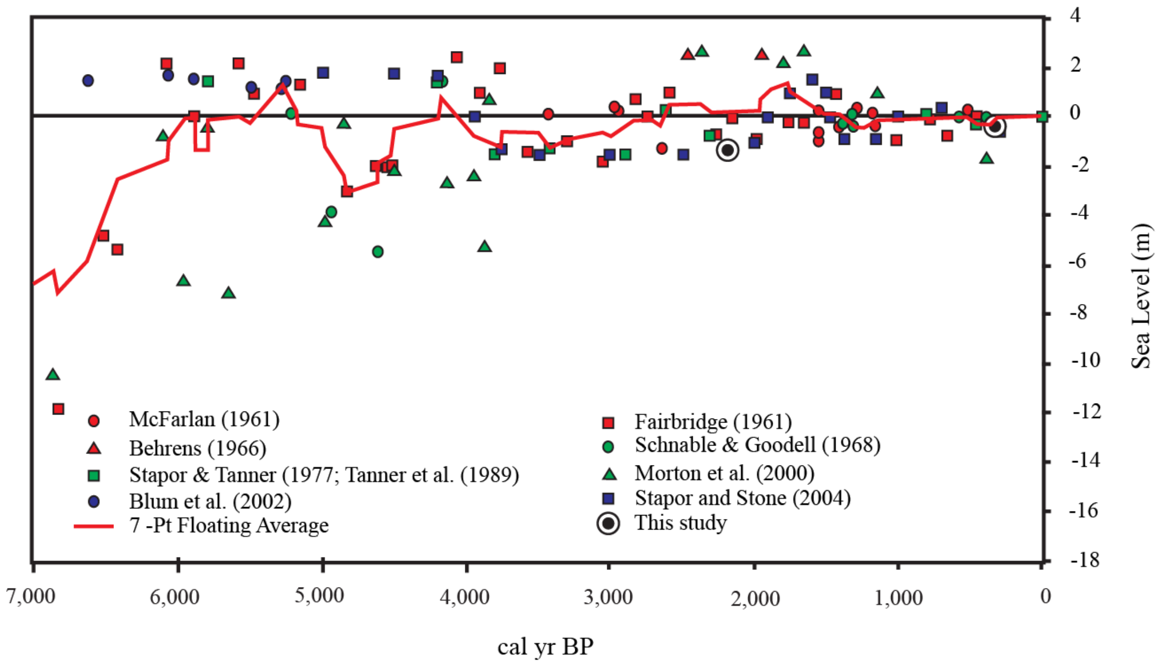

Figure 14. Gulf of Mexico younger data comprising dated sample sets collected onshore and offshore from the present shore line. Solid curve represents a 7-point floating average, which has been fitted to the calibrated age data set (Donoghue, J.F. and Balsillie, J.H., 2004)

Core	Interval (cm)	Sample material	MEASURED AGE (yrs B.P)	$^{13}C/^{12}C$	CONVENTIONAL AGE (yrs B.P)	2 SIGMA CALIBRATION
C1	116 - 115	Twigs	25,990±130	-24.6	26,000±130	
	331 - 330	Organic Sand	29,640±170	-22.1	29,690±170	
C2	59 - 58	Twigs	340±30	-26.1	320±30	Cal AD 1,470 to 1,650 (Cal BP 480 to 300)
	150 - 149	Organic Sand	2,180±30	-13.8	2,360±30	Cal BC 490 to 460 (Cal BP 2,440 to 2,410)/Cal BC 420 to 390 (Cal BP 2,370 to 2,340)
	302 - 301	Twigs	12,760±60	-24.6	12,770±60	Cal BC 13,250 to 13,000 (Cal BP 15,200 to 14,950)
	388 - 387	Twigs	NA	-25.4	>43,500	

Table 1. Radiocarbon dates.